THE UNIVERSITY OF TEXAS

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SEPTEMBER 1, 1960

# Aspects of the Geology of Texas: A Symposium

FRANK B. CONSELMAN; JOSEPH D. MARTINEZ, EDWIN H. STATHAM, AND LYNN G. HOWELL; HENRY F. NELSON; JOHN H. NICHOLSON; JAMES LEE WILSON AND O. P. MAJEWSKE; AND ADDISON YOUNG

IN CO-OPERATION WITH DEPARTMENT OF GEOLOGY, THE UNIVERSITY OF TEXAS

BUREAU OF ECONOMIC GEOLOGY

THE UNIVERSITY OF TEXAS, AUSTIN

JOHN T. LONSDALE, *Director*

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*The benefits of education and of useful knowledge, generally diffused through a community, are essential to the preservation of a free government.*

Sam Houston

*Cultivated mind is the guardian genius of Democracy, and while guided and controlled by virtue, the noblest attribute of man. It is the only dictator that freemen acknowledge, and the only security which freemen desire.*

Mirabeau B. Lamar

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# Foreword

The papers contained in this volume were presented at The University of Texas October 31, 1958 in a conference on "Aspects of the Geology of Texas." This conference was a part of the observance of the 75th Year of The University of Texas and was sponsored jointly by the Bureau of Economic Geology and the Department of Geology. One paper, "Deposition and Alteration of the Edwards Limestone, Central Texas," by Henry F. Nelson, is included in abstract form only, because it was published separately, as The University of Texas Publication No. 5905, to accommodate a field trip at the annual meeting of the American Association of Petroleum Geologists in Dallas in March 1959.

For this symposium papers were sought presenting new concepts or techniques in geology or new information on little known parts of Texas. It is hoped that these papers will stimulate advanced thinking on the geological problems of Texas which must be solved if the State is to retain its prominent position as a producer of petroleum, natural gas, and industrial rocks and minerals upon which its economy so largely depends.

JOHN T. LONSDALE, *Director*
Bureau of Economic Geology

SAMUEL P. ELLISON, *Chairman*
Department of Geology

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# Pennsylvanian Reef Patterns in West-Central Texas

**FRANK B. CONSELMAN**[1]

## ABSTRACT

The recognition of reefs has been based on many different criteria, but the reef concept is basically a very simple one. In actual practice, a reef is identified by its form and may be defined as a carbonate body whose upper surface is markedly convex, as a result of predominantly organic marine sedimentary processes.

West-Central Texas Pennsylvanian reefs may be classified into five categories: ridge reefs, "button" or round reefs (bioherms), chain or cluster reefs, atolls, and irregular or composite reefs. Examples of these patterns are the Jameson reef (ridge), Double Mountain reef ("button" or round), the reef system south of Merkel, Taylor County (chain or cluster), Horseshoe atoll (atoll), Round Top Canyon reef (irregular or composite), and others.

The physical parameters of reefs are useful in the statistical approach to reef classification and recognition. By directly observed quantitative criteria, stratigraphic carbonate anomalies of reefoid character are classified as follows:

(1) If contour closure equals or exceeds 100 feet per mile of width, the deposit is a reef.

(2) If contour closure is 50 to 100 feet per mile of width, the deposit is probably a reef.

(3) If contour closure is less than 50 feet per mile of width, the deposit is a bank or biostrome.

The physical parameters may be presented in a convenient shorthand tabulation, for example, $6 \times 2/8$ @ 3, which describes a reef 6 miles long and 2 miles wide with 800 feet of relief and a bearing of 30 degrees.

## INTRODUCTION

The shelf sea sediments of the Pennsylvanian of West-Central Texas provide an excellent subsurface laboratory for the study of productive and non-productive reef deposits. These reefs are numerous, well preserved, and of economic importance as oil and gas reservoirs, in a region where development has been sufficiently intensive over a period of years to afford a wide variety of patterns and case histories. They are of further professional importance in that a large number of petroleum geologists have studied West-Central Texas reefs and have thus obtained useful standards of reference and experience for exploration in similar environments.

A sufficient amount of practical subsurface information is now at hand to make it possible to consider subsurface reefs statistically and empirically and to attempt to derive valid generalizations based on actual occurrences. Analysis of these ancient reefs provides a stimulating basis for comparison with the modern examples on which our theoretical considerations have been based.

## ENVIRONMENT OF DEPOSITION

The stratigraphy of the Pennsylvanian deposits of West-Central Texas is an intricately detailed and complex subject, which has been extensively treated in the literature and which will undoubtedly receive additional attention in future years. Only a generalized summary of the environ-

[1] Consulting Geologist, Abilene, Texas.

ments of reefing falls within the scope of this paper.

The area herein discussed has been arbitrarily bounded to the west by the Horseshoe atoll, described in detail by Myers, Stafford, and Burnside (1956) (fig. 1). It extends eastward to the longitude of an axis drawn from Baylor to Menard counties, beyond which the greater part of the Pennsylvanian section is predominantly clastic, with reefing restricted to local facies of the Canyon, "Caddo," and Marble Falls limestones. To the east, truncation is a regional factor across the Bend arch.

Pennsylvanian sediments have the general character of a wedge, thinning westward and thickening rapidly eastward into the Fort Worth basin. Older beds, of Bend-Atoka-Lampasas equivalence, appear to have been influenced by local structural factors, both pre-existent and contemporaneous, but the remainder of the Pennsylvanian, from Strawn upward through Canyon and Cisco time, seems to have been free from major diastrophic effects, although fluctuations in sea level and sedimentary interruptions were numerous.

The combination of ecologic factors during most of the Pennsylvanian appears to have been favorable for reef development in a broad shelf area of West-Central Texas. Apparently reef growth could and did take place anywhere within a region of more than 25,000 square miles, in one part of the Pennsylvanian section or another. In some places reefing began early in the Strawn and progressed with apparent continuity upward through the Canyon into the Cisco, without paying noticeable attention to time lines. In other areas, reefing was confined to a small area and a restricted portion of the section, the causal factors being intangible as far as present ability to determine them is concerned. Traditional concepts such as shore-lines, lagoons, seaward faces, basement structures, seem to have little or no practical application in most cases.

## TERMINOLOGY

Three centuries ago the English naturalist John Ray wrote, "He that useth many words for the explaining any subject, doth like the cuttle fish, hide himself for the most part in his own ink." The term "reef" is fundamentally a very simple one and should be readily understood. The writer does not doubt that the thousands of petroleum geologists who use it as a matter of routine have a perfectly clear and satisfactory mental picture of what a reef actually is, and with the exception of borderline instances, or of people who would probably be mixed up anyhow, the word carries no confusion in normal use. This understanding is also largely shared by management and by the operating segment of the industry—in fact, the word "reef" could not be replaced without causing more trouble than it would cure.

Nevertheless, various qualifications, modifications, and amendments to the term have been offered from time to time, primarily by workers fresh from the laboratory, or from the library, and many theoretical pre-conceptions have been carried into the field. Criteria for recognition have been proposed on petrographic, paleontologic, and ecologic grounds. An essential condition imposed by Lowenstam (1950), Cloud (1952), and others is the existence of a wave-resistant structure. However, a very broad implication of possible difference between modern reef ecology of the Pacific type, and that of ancient reefs, is contained in an important contribution by Teichert (1958). This paper describes in convincing detail the occurrence at depths well below wave base, of organic formations dominated by the coral *Lophelia,* which would otherwise be classed as reefs. Preferred depths are 600 to 900 feet, off the coast of Norway, with reports of *Lophelia* occurring as deep as 3,000 feet. Usual water temperatures range from 6° to 6.5° C., or about 43° F. Accord-


FIG. 1. Reference map of West-Central Texas showing distribution of productive reefs.

ing to Teichert, the fauna includes at least 120 species capable of contributing to the bulk and growth of these deep and cold organic masses. Their relative proportions would certainly class them as reefs under other conditions. In this case the waters are far deeper and colder than previously considered admissible in a modern reef environment and well below the depth at which resistance to wave action would be necessary. In considering applications of these revised concepts to ancient deposits, Teichert suggests that the bioherm in the Winchell formation of middle Canyon age in McCulloch County, Texas, described by Young and Rush (1956), may have been formed in "deep and cool waters close to or at a moderate distance from a shore." This possibility would presumably also apply to other Canyon biohermal deposits elsewhere in West-Central Texas.

For practical purposes, a reef may be defined as a carbonate body whose upper surface is markedly convex topographically, as a result of predominantly organic marine sedimentary processes (Conselman, 1954). This agrees generally with a condensed version of the reef definition of Wilson (1950). However, it is suggested herein that in actual practice a reef is identified as such, not because of the presence of characteristic faunal or floral assemblages nor of specific lithologic facies, but because of its *form.* Criteria for recognition are not paleontologic nor petrographic, as these are frequently duplicated in non-reef deposits. Reefs essentially are characterized by relative topographic relief, and the surest criterion is non-structural contour configuration of a carbonate mass, in any of numerous contour patterns.

As a matter of practical observation, West-Central Texas reef patterns appear to fall readily into the following five general categories: ridge reefs, "button" or round reefs (bioherms), chain or cluster reefs, atolls, and irregular or composite reefs. This classification is rather simple and unimpressive and uses no long or foreign words, but it seems to include almost every generally recognized reef type in the area.

The present reef vocabulary features such terms as "bioherms" and "biostromes," but these appear to be more euphonious than useful, as many reefs are neither bioherms nor biostromes. A *bioherm,* as originally defined (Cumings, 1930, 1932, and *in* Cumings and Shrock, 1928), is a dome-like or mound-like mass, and while the term "mound" is not precise, and "dome" implies hemisphericity, bioherms are now usually considered as having a more or less round plan or pattern and would fit in the "round reef" category. The term *biostrome* has been applied to tabular deposits or lenses that are flatter or less convex than a mound or dome, and thus a biostrome is customarily considered equivalent to a bank, rather than a reef, in ordinary usage. The prefix "bio" now appears to be superfluous, as limestone reefs and banks are almost invariably thought of as having an organic origin, and no contrasting terms are in current use. In any case, "bioherm" and "biostrome" fall far short of providing an adequate vocabulary for reef pattern classification, although they admittedly have a rather convincing technical sound, which is quite impressive to non-professional audiences. The writer suggests they be used sparingly, for maximum effect.

"Ridge reefs," as here used, would include any linear or elongated reef, such as barriers and fringing reefs. The term has the advantage of avoiding the necessity of establishing that anything is being barred or fringed, which would normally be an impossible burden of proof under subsurface conditions. "Table reefs" and "patch reefs" are not without their appeal as descriptive terms, although "table" suggests the adjective "tabular," which in turn implies a different shape than intended. Subsurface-wise, the recognition of table reefs and patch reefs would be doubtful, since tables and patches are not conventional contour patterns.

In approaching the matter of identifying reefs as such, it is desirable to start with an example that will fit any definition of reef, as foolproof as can be had. For this purpose the Jameson reef in northwestern Coke County is nominated (fig. 2). This is a reef, and it looks like a reef, and while its time equivalence is subject to minor dispute, there is no recorded instance of any competent authority questioning that it is a reef. It is elongated in a northeast-southwest direction, as are the majority of Pennsylvanian ridge reefs in West-Central Texas. This elongation disqualifies it as a "bioherm." It is approximately 6 miles long and 2 miles wide and has about 800 feet of relief. In detail, a number of features would be observable, including terracing, clastic aprons, and probably pinnacles, which are commonly noted in other examples.

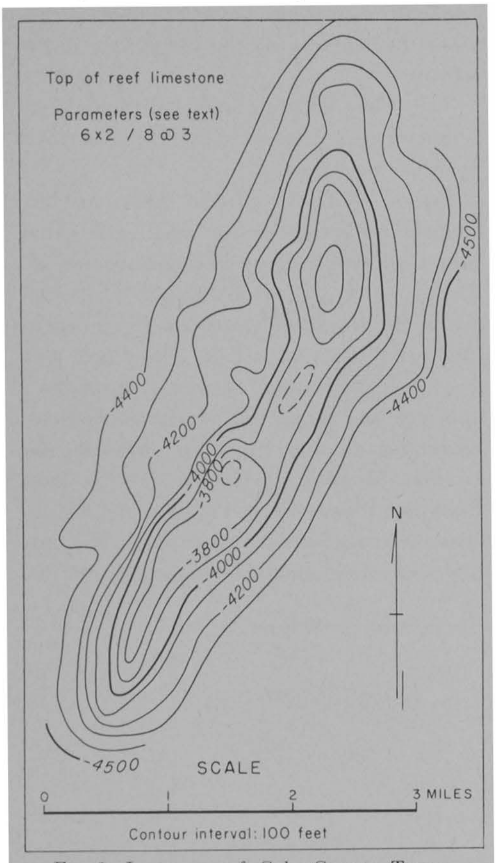


FIG. 2. Jameson reef, Coke County, Texas.

The Jameson reef figure includes the notation "6 x 2 / 8 @ 3," which is a shorthand tabulation of the important physical parameters of this reef body. These parameters represent one means of approaching reef classification from the statistical standpoint. The first two are length and width, to the nearest mile; the third is the relief in hundreds of feet (number of 100-foot contours); the last is the heading or direction of the long axis, in tens of degrees, true. These parameters are quite simple and easily derived; they can readily be refined to more precise measurements if the coarse figures do not provide sufficient detail. They are extremely useful in reef classification and recognition.

A ridge reef quite similar to Jameson in many respects is Claytonville, in Fisher County (fig. 3). It will be noted that the parameters reflect this similarity nicely.

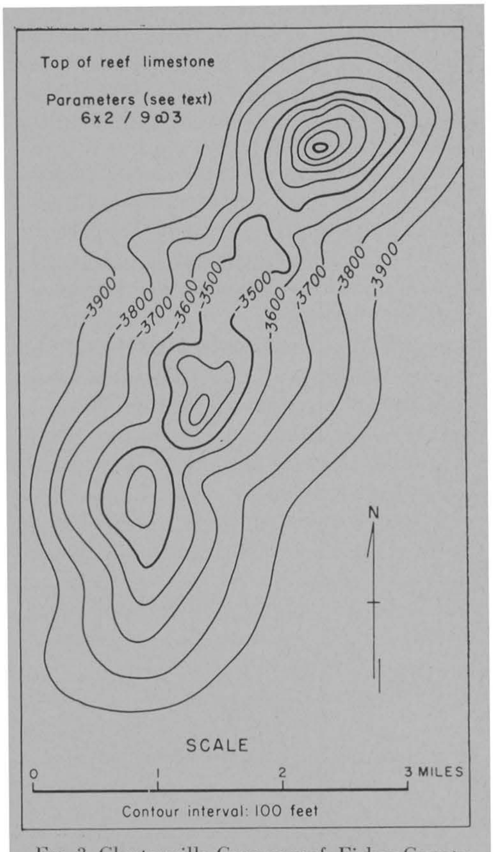


FIG. 3. Claytonville Canyon reef, Fisher County, Texas.

An interesting recent discovery is the I.A.B. reef (fig. 4), not yet fully developed. Even on the basis of incomplete control, it shapes up into a pattern quite similar to Jameson, 8 miles northwest. It took eleven years from the time of the discovery of Jameson to find this very promising reef reservoir, lurking only 8 miles away. This raises an interesting question as to how many more Jamesons, Claytonvilles, and I.A.B.'s are still hiding out within easy reach, not only in West-Central Texas but in similar sedimentary environments elsewhere.

The Millican reef (fig. 5), as shown in figure 1, is quite close geographically to both Jameson and I.A.B. but is anomalous to them in size and heading.

It will be noted that no attempt is made to establish an exact time equivalence for any of these reservoirs for the purpose of pattern comparison. The age of a reef may have economic as well as stratigraphic significance, as will be discussed later, but often millions of barrels of oil have been produced before a firm correlation can be established, if then. Usually very few embarrassing questions are asked about age if the reef is productive.

North Knox City (fig. 6) in Knox County is another type of ridge reef. It is noted for the fact that the discovery well was offset in opposite directions by dry holes, which seems to be carrying the ridge idea to extremes.

The Page Strawn reef in Schleicher County (fig. 7) has recently been described by Ellison (1957).

Round reefs, or button reefs, are extremely numerous in the area, and often they seem to spring up like mushrooms, although not nearly often enough. These undoubtedly would be classed as bioherms in the strict original sense. They are also known as "hickeys," "knobs," "pimples," and by other terms not nearly as classical. Examples in the literature include the Double Mountain (Conley, 1952) and Stamford (Van Siclen, 1957) reefs. All degrees of roundness are observed in the outlines of these small features. Many un-

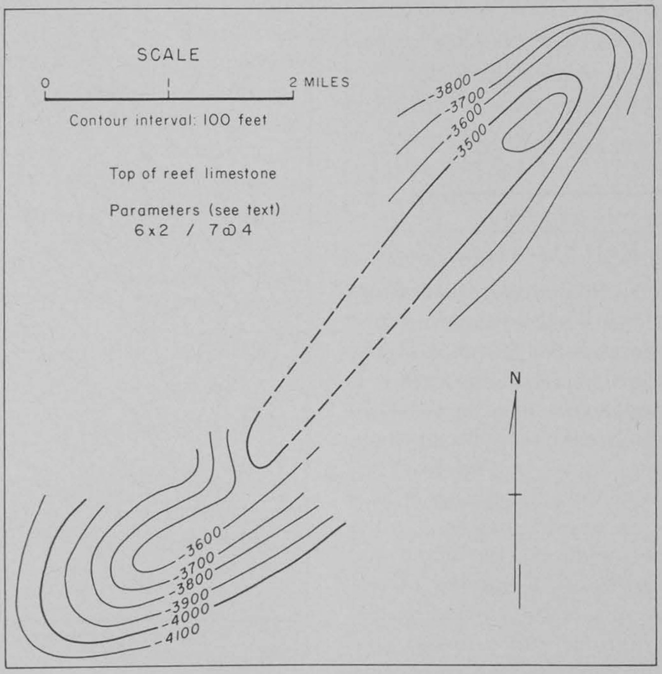


FIG. 4. I.A.B. reef, Coke County, Texas.

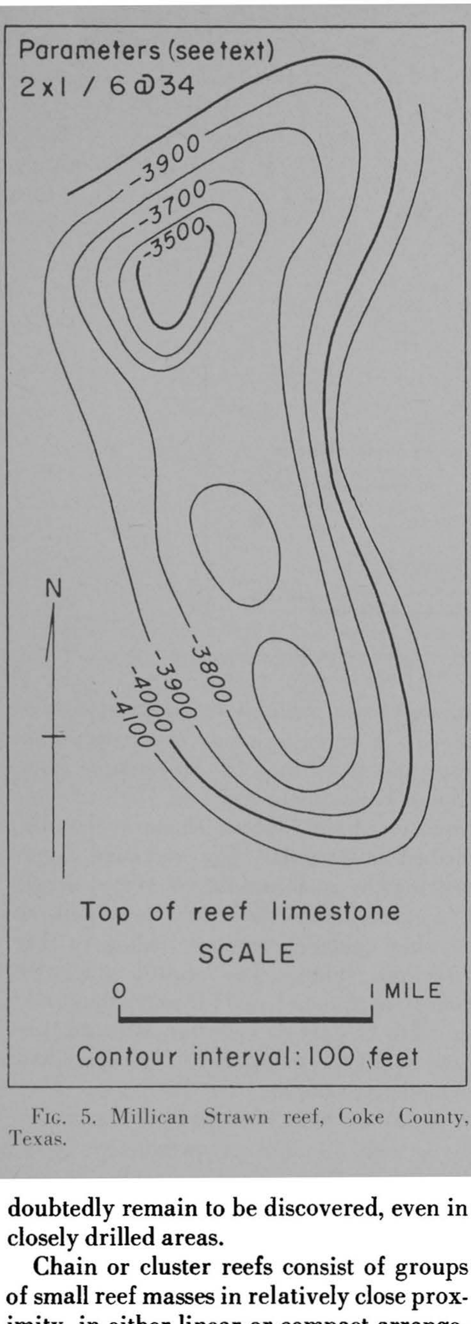


FIG. 5. Millican Strawn reef, Coke County, Texas.

doubtedly remain to be discovered, even in closely drilled areas.

Chain or cluster reefs consist of groups of small reef masses in relatively close proximity, in either linear or compact arrangements. Their occurrence is often intriguing and even tantalizing, because they inevitably suggest the existence of other undiscovered related reef elements nearby. Figure 8 shows two types of cluster reefs of Cisco age. The left-hand panel includes a segment of a chain reef system south of Merkel, in Taylor County, the chain being traceable for many miles as a double string of small producing pools. This chain may eventually tie in southward to the very similar features in northern Runnels County. The right-hand panel shows the cluster of small reefs in the Fennell area of Runnels County. A major chain in Sutton and Schleicher counties has components varying in pattern from ridges to knolls; this chain has been described by Rall and Rall (1958). The Griffin-Avoca-Ivy chain in northeastern Jones and northwestern Shackelford counties is well and favorably known from the commercial standpoint, and numerous other belts and clusters have been drilled.

Figure 9 shows two closely related reefs in northwestern Nolan County—Rowan & Hope and Rowan & Hope Northwest. The reef to the right is under the airport west of Sweetwater; the one to the left is a short distance northeast of Roscoe. These reefs were undoubtedly contemporaneous and have almost identical summit accordances. They would appear to be part of a larger system, but as pointers their message remains a bit ambiguous.

Atolls are a very picturesque type of reef and probably figure vividly in everyone's background conception of reef development. Unfortunately, relatively few bona fide atolls have been identified, the enormous Horseshoe atoll apparently having consumed almost the entire Pennsylvanian allotment for the area. However, the Miers Strawn reef in southeastern Sutton County appears to be part of an atoll system approximately 6 miles in diameter, as contoured by Nichols (1957), and the Miers gas field has the additional distinction of being perched directly on top of a pre-Paleozoic sea-mount. Possibly a number of our cluster reefs will prove to be parts of atolls when the missing perimeter segments are discovered.

Several reef complexes do not lend themselves readily to classification in the foregoing patterns, and for them Category 5, for irregular, composite, and miscel-

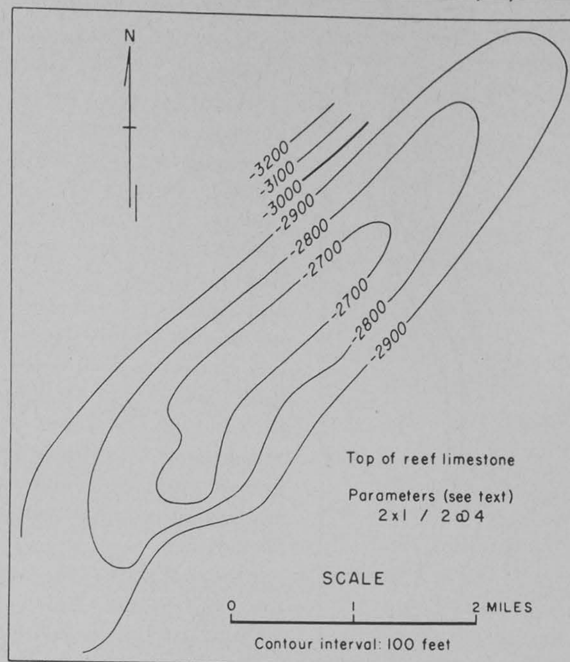


FIG. 6. North Knox City Canyon reef, Knox County, Texas.

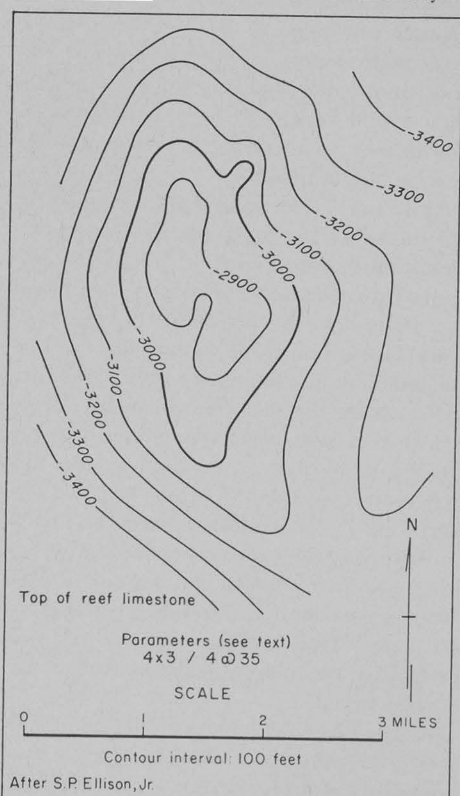


FIG. 7. Page Strawn reef, Schleicher County, Texas.

laneous types, is highly recommended, even though it suggests a sort of stratigraphic sweeping under the rug. An example is the Round Top Canyon reef (fig. 10) in Fisher County, whose contour shape is best described as irregular. The westward extension may be an accumulation of reef debris as a clastic apron, or it may be responsive to other contemporaneous reefing in that direction. Round Top would otherwise qualify as a round reef; it may eventually prove to be part of a cluster. It could perhaps qualify as a "patch reef" as described in modern examples.

So far we have been dealing almost entirely with oil- and gas-productive reefs. Attention is now invited to a really impressive reef mass lying east of Anson, in Jones County (fig. 11). This reef is 16 miles long and 7 miles wide and has over 800 feet of known relief. It lies in the heart of the producing reef area and has been penetrated by almost a hundred rotary and cable tool tests, yet not a single commercial "reef" well has been completed in it. There is no question of permeability—in fact, it appears to be water-logged throughout, and cable tool holes have reported one hole

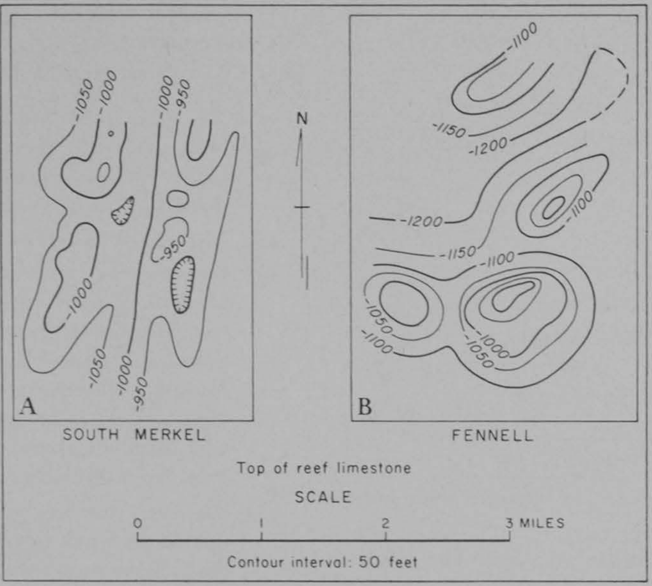


FIG. 8. South Merkel area, Taylor County, Texas, and Fennell area, Runnels County, Texas.

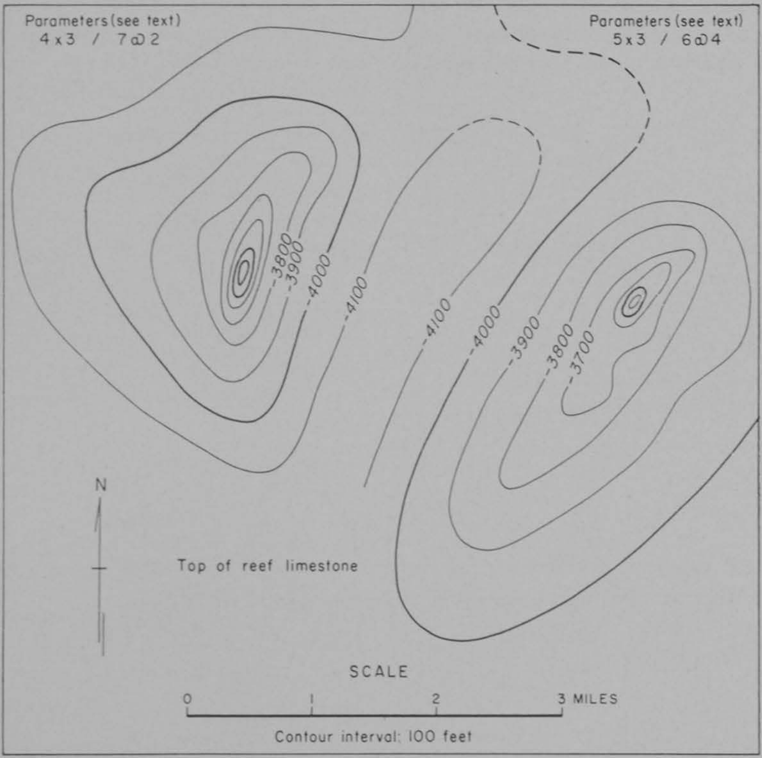


FIG. 9. Rowan & Hope and Rowan & Hope Northwest reefs, Nolan County, Texas.

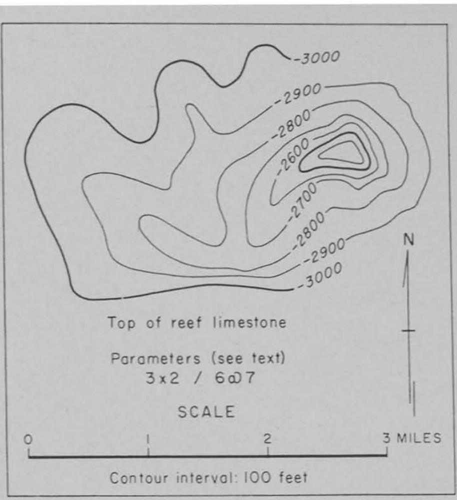


FIG. 10. Round Top Canyon reef, Fisher County, Texas.

full of water below another. The permeability of this East Anson reef is in fact suggestive of the results recently obtained in ground-water observations in the Caroline (McKee, 1958) and Marshall (Swartz, 1958) Islands, where tidal and thermal responses to oceanic conditions have been described and attributed to high permeability.

The question naturally arises as to why this reservoir apparently carries no oil. There are several possible explanations:

(1) The oil is there, but the reef has not been drilled at the right spot. Actually, dozens of wells have been drilled in the terrace area.

(2) The age of the reef—chiefly Middle Canyon—is unfavorable. Lower Canyon (Palo Pinto) reefs produce immediately to the north, and oil is found in Upper Canyon reefs and Middle Canyon sands in nearby areas. There is no apparent reason why Middle Canyon ecology should have been unfavorable here.

(3) The reservoir has been flushed, and the oil and gas have escaped, to a destination presently unknown.

Regardless of the reason, this reef certainly belies the idea that all reefs carry oil. Other large water-logged masses of similar characteristics have been known for years, including one submerged in central and western Taylor County, that may be the contemporary of this East Anson reef.

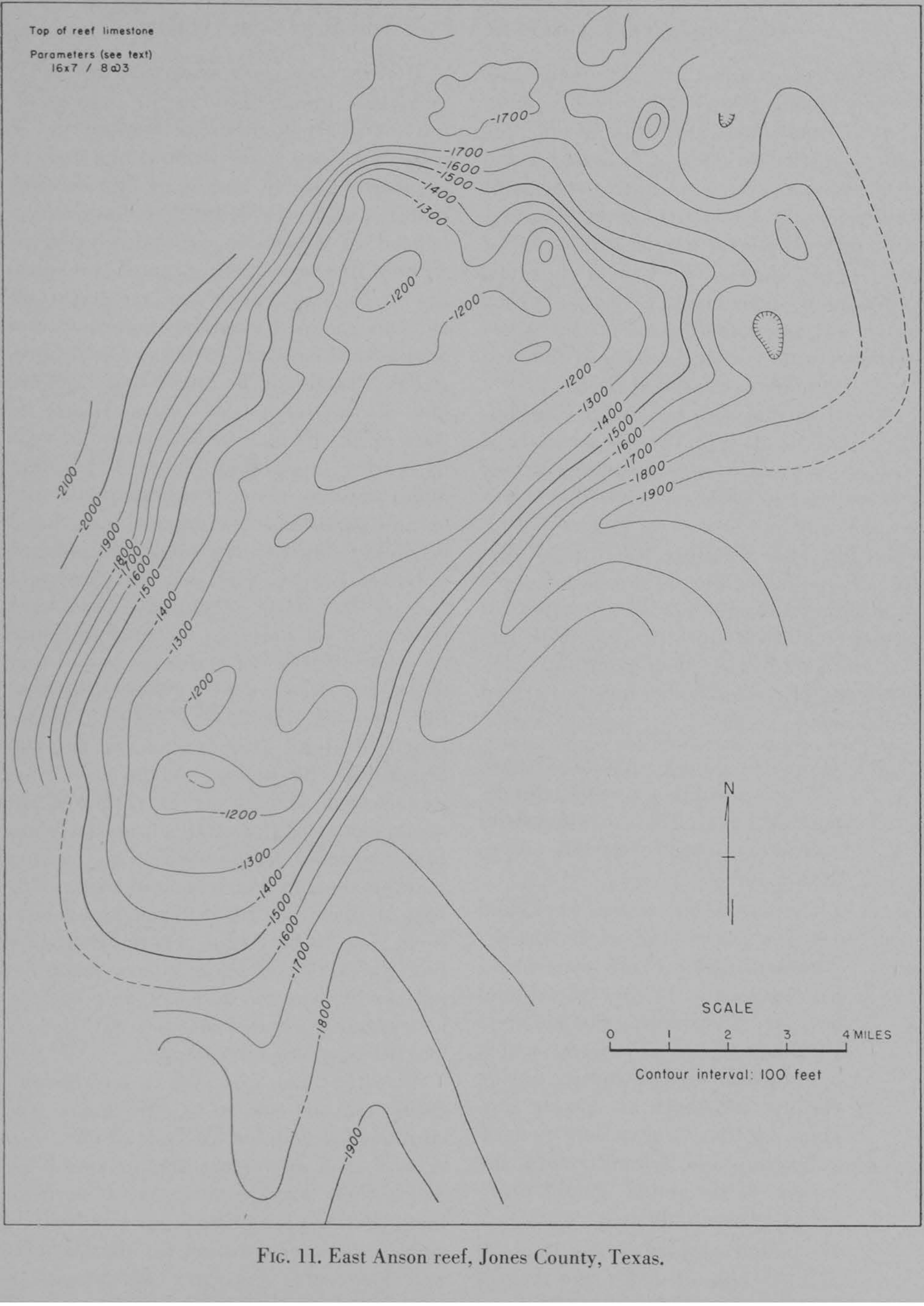


FIG. 11. East Anson reef, Jones County, Texas.

# QUANTITATIVE CRITERIA FOR REEF RECOGNITION

In reviewing statistically the many subsurface reefs and reef-like masses in West-Central Texas, there appear to be sufficient data to justify an attempt to define a reef in terms of specific quantitative criteria or parameters, as a matter of direct observation, rather than on purely qualitative or theoretical grounds. This type of empirical definition would conform to the established usage and understanding of the many geologists familiar with the reefs of the area on an intimate professional basis.

Assuming that we have under consideration a sedimentary, carbonate anomaly of reefoid character, and assuming further that we have satisfied ourselves that stratigraphic and not structural factors are responsible, and assuming still further that we have not fallen into the elementary error of mistaking normal marine gradation for reefing—in other words, if *qualitative* factors are not adverse—then *quantitative* expressions for reefoid bodies may be derived as follows:

(1) If contour closure equals or exceeds 100 feet per mile of width, the deposit is a reef. This corresponds to a gradient of approximately one in fifty.

(2) If contour closure is between 50 and 100 feet per mile of width, the deposit is probably a reef, usage being less unanimous in this transitional category. Considering the post-depositional history of reefs, which may include such disruptive or distorting influences as marine erosion, solution, compaction, re-crystallization, and dolomitization, the benefit of the doubt should probably be given to the reef.

(3) If contour closure is less than 50 feet per mile of width, the deposit is more properly termed a bank, or biostrome if preferred, and should be further checked to make sure structural influences are not primarily responsible.

It may seem somewhat unorthodox to emphasize quantitative rather than qualitative criteria in reef classification and to call a limestone a reef without first making absolutely sure by means of thin sections that it is made of calcilutite, or calcarenite, instead of calsomine or chalcopyrite, or without first ascertaining that it is a seething mass of coralline or algal material rather than calcified dinosaur bones, animal crackers, or broken crockery. The fact remains, and should be faced, that reefs are and always have been characterized by their *form*, and that they are a specific type of marine topographic form. Given that external form under Pennsylvanian subsurface conditions, the internal faunal and lithologic content may safely be assumed to be satisfactory. Even the "wave-resistant framework" must necessarily have been present, if its presence is required, since the existing relief of the mass proves that the waves, if any, must have been successfully resisted. The lowly *Lophelia*, whose frigid, if rigid, little framework lies 800 feet below the surface of the cold Norwegian seas, should not be forgotten. In actual practice, almost all of our reefs are first identified by the effect of reef-top markers on subsurface contour maps. This may be deplored, but it works, and it does so in time to be useful. There are always limitations, of course, as to how much detail can be expected under 40-acre and 80-acre spacing, whether one uses electric logs or a petrographic microscope.

The statistical approach to reef classification may be refined to offer many possibilities for detailed analysis of these reservoirs, with interesting implications as regards their local and regional environment. Pattern parameters are particularly significant in exploration for similar reefs and may offer corollary advantages in studying associated flank sediments, fluid migration, and post-depositional history. It also offers an important opportunity for improved terminology having quantitative definition.

West-Central Texas is certainly not unique as regards its Pennsylvanian shelf deposits, and data derived in this area should be useful elsewhere, and possibly also in other parts of the subsurface section in which reefing occurs. In the meantime, we have much still to learn and many undiscovered reefs still to find close at home, when market demand once more makes oil a valuable commodity.

## REFERENCES


CLOUD, P. E., JR. (1952) Facies relationships of organic reefs: Bull. Amer. Assoc. Petrol. Geol., vol. 36, pp. 2125–2149.

CONLEY, J. N. (1952) Double Mountain pool, Stonewall County, Texas: Abilene Geol. Soc. Contributions 1952, pp. 17–20 and figs.

CONSELMAN, F. B. (1954) Oil and gas production from carbonate reservoirs: Amer. Assoc. Petrol. Geol., annual meeting, 1954, oral presentation.

————, and CONLEY, J. N. (1949) Limestone development in the Pennsylvanian of the eastern Permian basin: Amer. Assoc. Petrol. Geol., annual meeting, 1949, oral presentation.

CUMINGS, E. R. (1930) List of species from the New Corydon, Kokomo, and Kenneth formations of Indiana, and from reefs in the Mississinewa and Liston Creek formations: Proc. Indiana Acad. Sci., vol. 39, pp. 204–211.

———— (1932) Reefs or bioherms?: Bull. Geol. Soc. Amer., vol. 43, pp. 331–352.

————, and SHROCK, R. R. (1928) Niagaran coral reefs of Indiana and adjacent states and their stratigraphic relations: Bull. Geol. Soc. Amer., vol. 39, pp. 579–620.

ELLISON, S. P., JR. (1957) Page field, Schleicher County, Texas, *in* Occurrence of oil and gas in west Texas: Univ. Texas Pub. 5716, pp. 247–252.

LOWENSTAM, H. A. (1950) Niagaran reefs of the Great Lakes area: Jour. Geol., vol. 58, pp. 430–487.

MCKEE, E. D. (1958) Geology of Kapingamarani atoll, Caroline Islands: Bull. Geol. Soc. Amer., vol. 69, pp. 241–278.

MYERS, D. A., STAFFORD, P. T., and BURNSIDE, R. J. (1956) Geology of the late Paleozoic Horseshoe atoll in west Texas: Univ. Texas Pub. 5607, 113 pp.

NICHOLS, J. M. (1957) Miers field, Sutton County, Texas, *in* Occurrence of oil and gas in west Texas: Univ. Texas Pub. 5716, pp. 214–217.

RALL, R. W., and RALL, E. P. (1958) Pennsylvanian subsurface geology of Sutton and Schleicher counties, Texas: Bull. Amer. Assoc. Petrol. Geol., vol. 42, pp. 839–870.

SWARTZ, J. H. (1958) Geothermal measurements on Eniwetok and Bikini atolls, *in* Bikini and nearby atolls, Marshall Islands: U. S. Geol. Survey Prof. Paper 260-U, pp. 711–741.

TEICHERT, CURT (1958) Cold- and deep-water coral banks: Bull. Amer. Assoc. Petrol. Geol., vol. 42, pp. 1064–1082.

VAN SICLEN, D. C. (1957) Organic reefs of Pennsylvanian age in Haskell County, Texas: Geophysics, vol. 22, pp. 610–629.

WILSON, W. B. (1950) Reef definition: Bull. Amer. Assoc. Petrol. Geol., vol. 34, p. 181.

YOUNG, K. P., and RUSH, R. W. (1956) Shape and deposition of small Pennsylvanian bioherm, McCulloch County, Texas: Bull. Amer. Assoc. Petrol. Geol., vol. 40, pp. 1988–1994.

# A Review of Paleomagnetic Studies of Some Texas Rocks

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## ABSTRACT

The various mechanisms by which rocks acquire a permanent or remanent magnetization are briefly reviewed. The positions of Mesozoic, Paleozoic, and Precambrian poles deduced from measurements of remanent magnetism on a world-wide basis by various workers is summarized. A short discussion covers the lines of reasoning which suggest that the ancient magnetic pole locations are generally coincident with the time equivalent geographic poles if a sufficiently long time interval has been included in the sampling.

The results of paleomagnetic studies at the Humble Research Center of a large number of varied rock types occurring in Texas are summarized and in certain instances discussed in some detail. The bulk of these measurements involve sedimentary rocks, but many igneous rocks and some metamorphics have also been investigated. While some data previously published are reviewed, this paper includes much material not yet published.

The new work reported on here includes paleomagnetic studies made of Tertiary volcanics in the Big Bend National Park. The purpose of this work was to test (and use where possible) paleomagnetic data for correlation purposes. Drs. John T. Lonsdale and Ross A. Maxwell of the Bureau of Economic Geology of The University of Texas were responsible for relating the sampling to the geology of the area. Most of the field work was conducted jointly with Dr. Maxwell and to some extent with Dr. Lonsdale. They have been engaged for some time in mapping this area, but their work is as yet largely unpublished. Without their guidance, this kind of study would not have been significant.

## INTRODUCTION

Paleomagnetic studies have been conducted for a number of years in the research laboratory of the Humble Oil & Refining Company. In the course of this work, a wide variety of Texas rocks have been examined. The lithologic types studied have included the three broad categories: sedimentary rocks, igneous rocks, and metamorphic rocks. A fair, although incomplete, sampling has been made of rocks ranging in age from Precambrian to Recent.

Some of the results obtained from these studies, which seemed to have particular significance, have already been published. However, most of the data have not been published prior to this time. This paper attempts to summarize selected typical data obtained from our studies of Texas rocks. The coverage is by no means encyclopedic, since many of the results are quite similar and to include everything would result in a very long and monotonous review.

First of all, the mechanisms by which rocks become magnetized are briefly reviewed. Most rocks, either igneous, metamorphic, or sedimentary, are at least weakly magnetic. This magnetism that is retained after removal of a magnetizing field is termed remanent magnetism. In

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theory, igneous rocks, both extrusive and intrusive, should acquire a remanent magnetism in the direction of the magnetic field in which they are cooled from a molten condition, or in some instances in a reverse direction to that field. The temperature at which this remanence is acquired is called the Curie point. It has been found by various workers in the laboratory that igneous rocks, after being heated, are magnetized in the direction of the applied magnetic field upon being cooled through the Curie point. Measurements on lava flows of recent years have also revealed magnetizations in the direction of the earth's field as known at the time of the flow. It should be mentioned that many igneous rocks have been found to be magnetized inconsistently in direction for various samples. Suggested explanations have been: (1) movement or flow in the rock after cooling through the Curie point and (2) remagnetization. On the other hand, many igneous rocks are consistently magnetized and in directions frequently quite different from the present earth's field. In fact, a succession of lava flows may exhibit a reversal of magnetization. About half of the Cenozoic lava flows (Handbuch der Physik, 1956) show a magnetization direction reverse to the usual direction. The manner in which sedimentary rocks have acquired their remanent magnetism is not so clearly established. One mechanism that has been proposed is that the settling of tiny magnetic particles in quiet water would result in their orientation by the earth's field. Another, which would apply in the case of chemical or oxidized sediments, is that the rock becomes magnetized in the direction of the earth's field when crystallization of the magnetic minerals from a solution or gel or recrystallization occurs. Another less intense, less stable, and hence less important mechanism of magnetization is termed isothermal. This mechanism involves a magnetism resulting simply from exposure to a magnetic field with no change in temperature.

The value of paleomagnetic measurements lies in the possibility that they may be used to trace the history and changes of the earth's magnetic field through geologic time. Of greater importance, granted certain assumptions are correct, is the possibility of establishing whether or not polar wandering or continental drift has ever occurred and to trace such possible movements. Both theoretical and, to some extent, observational evidence tend to confirm the hypothesis that when averaged for periods of time (several thousands of years) the earth's rotational or geographic and magnetic axes have been generally coincident. The dynamo theories of earth magnetism of Elsasser (1946, 1947) and Bullard (1949) suggest such a coincidence. In the case of rocks (principally lavas) of Miocene age or younger, Hospers (1955) has concluded that magnetic pole locations, made from averages of samples taken over time intervals sufficient to reduce secular deviations, are very close to the present geographic pole. Thus, he concludes that the geographic poles have not moved appreciably since the end of the Eocene.

Runcorn (1955a), Graham (1949), and others have found excellent evidence that the magnetic poles and very likely the geographic poles have moved and that the pole locations during Paleozoic time were quite different than they are today. Figure 12 is a map of the mean inferred geographic pole locations for the Paleozoic published by Howell and Martinez (1957) from their determinations and those of others. All of the data upon which these pole locations are based were obtained from sedimentary rocks. This map suggests that there was a progressive movement of the poles during Paleozoic time. Runcorn (1956a) postulated a relative movement between North America and Europe of about 20 degrees based on some of the data used in the preparation of this map. Other workers have invoked continental drift to

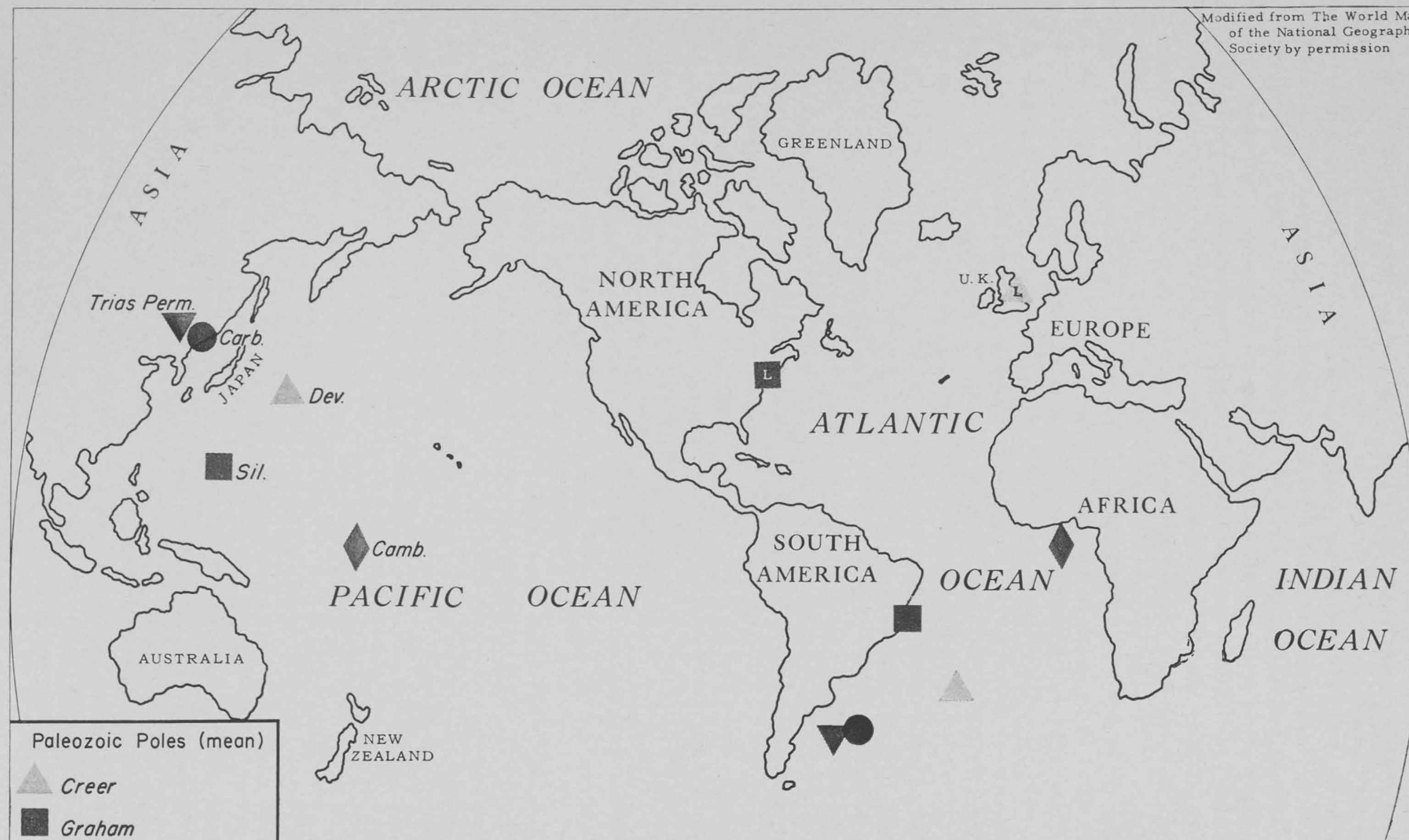


FIG. 12. Map of Paleozoic poles (after Howell and Martinez, 1957). Used by permission of "Geophysics."

explain divergent data obtained in land masses in the southern hemisphere.

Superimposed on this movement of the polar areas and/or relative drift of large continental masses is an apparent repeated reversal in sense of the earth's magnetic field that has already been mentioned. There is evidence which favors this hypothesis. On the other hand, several theoretical explanations, not requiring a reversal of the earth's field, have been offered to account for the commonly found inverse magnetization of rocks (Néel, 1951, 1955). Nagata (1952) actually verified one of these explanations experimentally. At present, the evidence appears to favor actual reversals of the magnetic poles of the earth.

We would next like to review the data obtained from paleomagnetic measurements of Texas rocks upon which our map of polar wandering is partially based as well as other data both published and unpublished.

Figure 13 shows locations of Precambrian poles determined by various workers. Our pole location at 49° N Latitude and 175° W Longitude (Howell, Martinez, and Statham, 1958) was based on measurements of the Hazel formation, a red-bed sequence, of Precambrian age cropping out in Culberson and Hudspeth counties of west Texas. It is interesting to note that a large number of paleomagnetic studies have been made with red beds, since hematite, which is thought to be the principal magnetic component of these beds, has a high magnetic coercive force. This means that the rock tends to retain its early magnetization. Our determination resulted from measurements of 15 samples of flat-lying beds from five locations scattered over about 2 miles. In order to test the stability of the remanent magnetism, 39 samples were obtained from nine locations in which the beds were dipping —in some instances, as high as 85°; these locations covered a distance of about 20 miles. Graham (1949) described basic tests for determining the stability of remanent magnetism with respect to geologic time. One of his tests involves measuring the directions of remanent magnetism in a folded bed or beds. If the magnetic vectors plotted on a system of space coordinates are scattered but brought back into congruence or their scatter reduced by rotating the beds back to a horizontal position, with a corresponding rotation of the magnetic vectors, then the remanent magnetism must have been stable at least since the time of folding. In the case of the Hazel sandstone, this test was inconclusive. The data from the dipping beds were considerably more scattered from that of the flat-lying beds. Figure 14 is a Schmidt net plot showing circles of confidence on the 95% level plotted for three groups of data. Circle A represents the data from the flat-lying beds, circle B those from the dipping beds plotted with reference to a system of space coordinates, and circle C those from the dipping beds plotted with reference to a system of coordinates tied into the attitude of the bed. If circle A had fallen within the area of circle B, then instability would have been indicated; if it had fallen within circle C, then stability would have been indicated. These results, as they stand, are inconclusive. However, the fact that the data from the flat-lying beds are consistent and they are magnetized in a direction quite different than the direction of the earth's field at the present time is evidence favoring stability for the flat-lying beds, although the samples show slight instability in the laboratory. Since some of the samples are from thin lenses of sandstone in thick sequences of conglomerates, a chemical rather than a detrital mechanism of magnetization is suggested. Figure 15 is a view of a typical sandstone lens of this sort; however, the lenses sampled were larger.

The Barnett formation of Mississippian age which crops out in the Llano uplift area of central Texas was sampled at the locations shown in figure 16. The results have been discussed in detail by Martinez and Howell (1956) and Howell and Martinez (1957). Most of the samples obtained were calcareous concretions, as shown in figure 17; however, a few were limestone and shale. The samples from one location were magnetized in a direction opposite to that of the samples from other localities. The pole locations determined from these directions of magnetization are shown in figure 18 as well as pole locations determined from Carboniferous rocks in England by Belshe (Runcorn, 1955b) and pole locations determined from Pennsylvanian rocks in the Arizona area by Runcorn


PRECAMBRIAN POLES

| AUTHORITY | FORMATION | AGE | GEOGRAPHICAL LOCATION OF SAMPLES |
|---|---|---|---|
| Irving's data by Runcorn | Torridonian | Late Precambrian | Great Britain |
| Creer's data by Runcorn | Longmyndian | Late Precambrian | Great Britain |
| R Runcorn | Hakatai shale | Late Precambrian | Grand Canyon, USA |
| D Doell | Hakatai shale | Late Precambrian | Grand Canyon, USA |
| C Creer, Irving, and Runcorn | Diabaig group Torridonian | Earlier Precambrian | Northwest Scotland |
| G Graham | Diabase dikes | Earlier Precambrian | Michigan, USA |
| H Humble | Hazel sandstone | Precambrian | Van Horn, USA |

FIG. 13. Map of Precambrian pole locations.

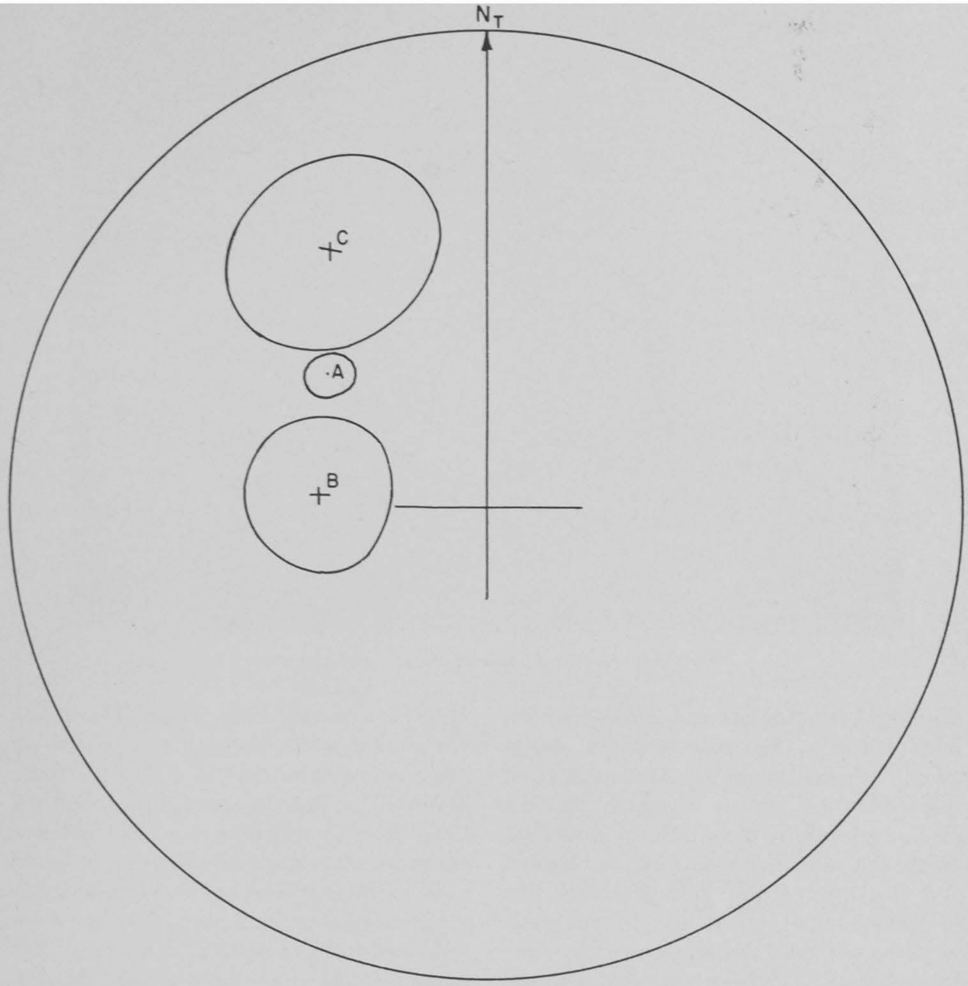


Fig. 14. Circles of confidence, plotted in lower hemisphere, for data obtained from the Hazel formation.

(1955b). The "starred" locations were calculated from the set of Barnett data obtained from the larger group of uniformly magnetized samples, while the plain circles represent pole calculations based on the data from the smaller reversely magnetized group of samples. An equator has been drawn on this map corresponding to our "starred" poles.

The Point Peak member of the Wilberns formation of Cambrian age which crops out in the Llano uplift area has also been sampled and the results published by Howell and Martinez (1957). The sample locations are also shown in figure 16. Most of the samples are siltstone. Figure 19 is a Schmidt net plot of magnetic vector directions determined for this group of samples. All points, except those that are dotted, fall in the lower hemisphere. All of the vectors from tilted beds have been rotated in a direction and amount consistent with the rotation required to remove the tilt of the beds. The data show more scatter than those for the Barnett. However, assuming the points nearer the present field indicate less stability than those more distant, a representative vector has been chosen near

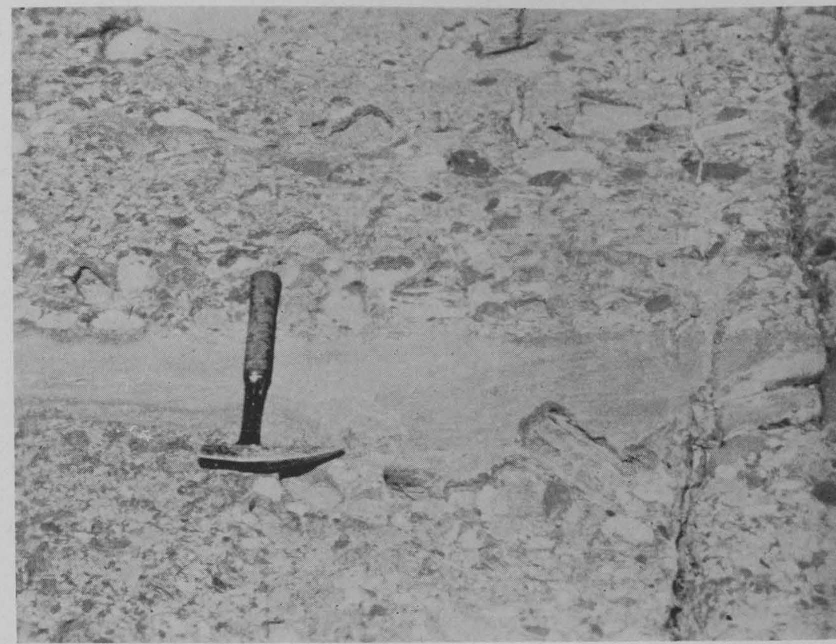

FIG. 15. Lens of sandstone in Hazel conglomerate.

the head of the points and marked with a multiple circle. The pole locations, designated by gray triangles, computed from this vector are shown in figure 20. The similar symbols with an L in the center indicate the sampling area location. Howell and Martinez (1957) have discussed this in greater detail, but it may be advisable to point out that measurements from the Keweenawan had been included because of a possibility that it might be of Cambrian age. Also, Humble poles designated by black squares were determined from Cambrian rocks that may have been remagnetized near the end of the Paleozoic.

While the following data have not been used to predict pole locations, principally because of lack of evidence for stability of remanent magnetism, these results are interesting as examples of some of the ramifications which appear in this field.

Figure 21 is a lower hemisphere projection on a Schmidt net of measurements of remanent magnetism of a large number of plugs obtained from two recumbent penecontemporaneous folds of siltstone in the Smithwick shale cropping out at Mormon Mills near Marble Falls, Texas. These data are plotted with reference to a system of space coordinates tied to a level surface; however, the beds involved dip 21 degrees. Regardless of which part of the fold was examined, the magnetic vector was found to be to the southeast and down, indicating magnetization after folding that has since remained stable. Figure 22 is a view of this folded sample after removal from the outcrop. Thirteen additional samples were obtained from five locations from the east, northeast, and west sides of the Llano uplift. Figure 23 shows the direction of magnetization of these samples as well as that of the folded sample plotted with reference to a system of coordinates tied into the bedding plane. The direction of the present field as well as that of the dipole field is also shown. The latter is the field that would result if the earth's magnetization is due to a simple dipole at the earth's center, the axis of which is congruent with the spin axis. A great circle has been drawn through the point indicating the direction of this dipole field and very generally following a zone of heavier concentration of plotted points. There is some indication of a drift of the directions of magnetization of the various samples along this great circle direction toward the present dipole field. The fact that many of the data are from dipping beds makes this approximation somewhat unreliable since the great circle route along which each vector would tend to move would be established by its position in space. At any rate, if an average vector were selected at the head of points as in the case of the Point Peak member of the Wilberns, the corresponding location of the North Pole would be 22° S Latitude and 56° W Longitude. The South Pole would be at 22° N Latitude and 124° E Longitude. This pole location, while in the general area of other determinations of Paleozoic poles, does not lie in its proper position along the proposed path of polar movement. As Runcorn (1956a) has pointed out previously, in the case of unstable magnetizations which follow a great circle through the dipole point, the longitude determinations should be more reliable than the latitude. All points on the great circle

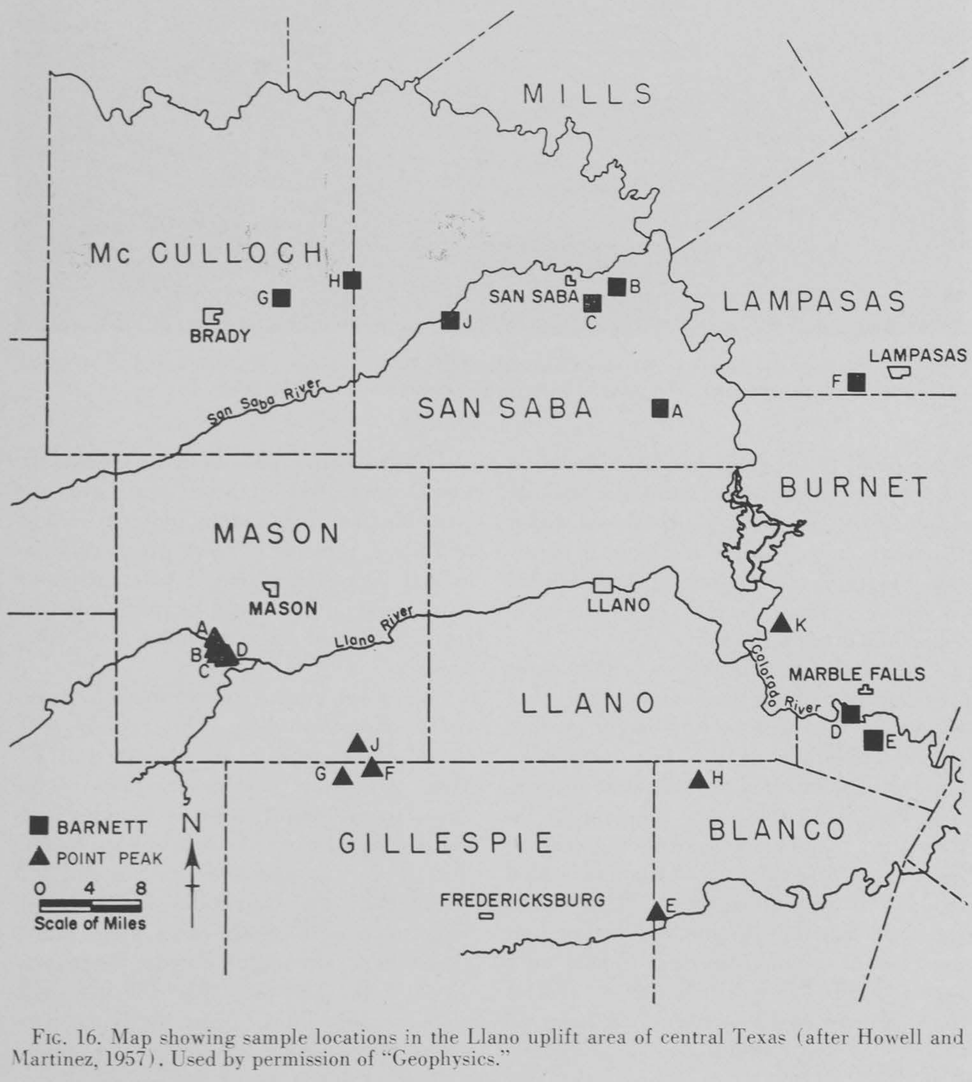


FIG. 16. Map showing sample locations in the Llano uplift area of central Texas (after Howell and Martinez, 1957). Used by permission of "Geophysics."

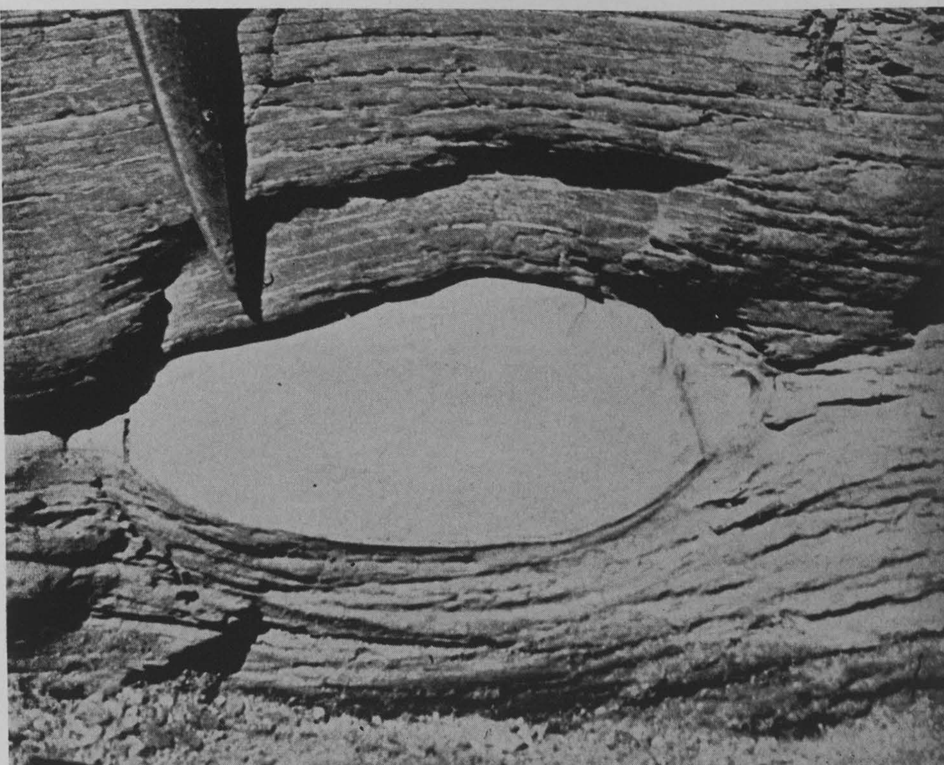

FIG. 17. Typical concretion in Barnett formation of Mississippian age (after Howell and Martinez, 1957; used with permission). The point of a geologist's hammer is shown for scale.

correspond to the same longitude location of the pole. On the other hand, the latitude depends on the head point chosen along the great circle. This point may be in error. Our longitude determination is somewhat in better agreement with other observations than is our latitude.

One of the real problems which enter into a study of the paleomagnetism of a suite of surface samples is to decide to what extent oxidation of the ferromagnesian minerals has affected the magnetic properties. Figure 24 shows the increase in intensity of magnetization with increase in the degree of oxidation of an unoriented boulder of Strawn sandstone. This would certainly lead one to the conclusion that the time of magnetization is indeed very recent. Figure 25 is a Schmidt net plot of the magnetic vectors determined from 18 samples of Bell Canyon sandstone of Permian age which crops out in the Guadalupe and Delaware Mountains of west Texas and New Mexico. These data indicate remagnetization, possibly by some process similar to that suggested above. A pole calculated from these points would be quite different from published Permian poles from other areas.

Somewhat similar data to those from the Bell Canyon formation, although more scattered, are shown in figures 26 and 27. These are lower hemisphere plots of the magnetic vectors of suites of samples from the Bedias member of the Wellborn formation of Eocene age and the Soledad tuff member of the Catahoula formation of Miocene age. In both cases these points cluster near, although not upon, the present dipole field. The principal difficulty with plots of this kind, from flat-lying beds where the vectors group so near the present

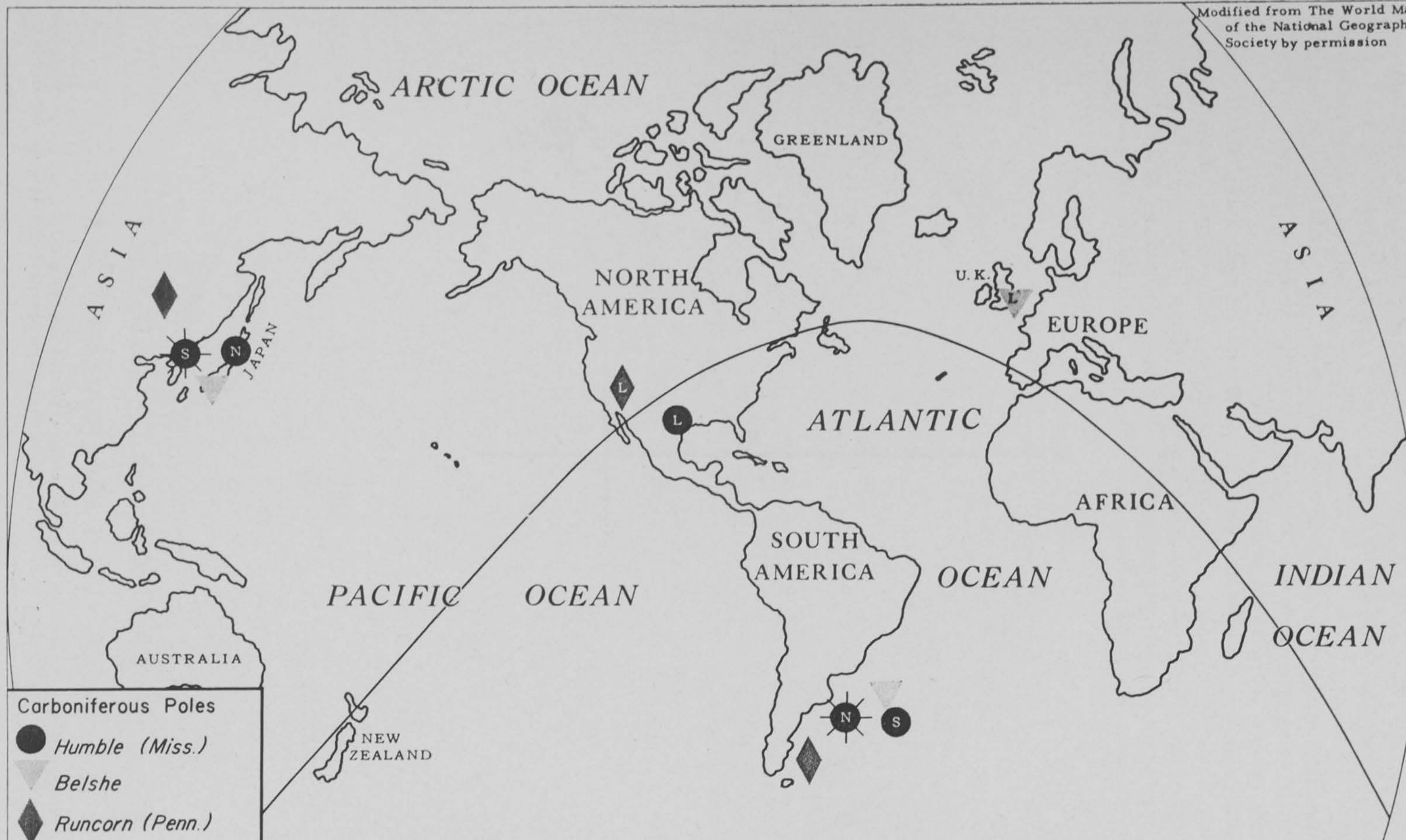


FIG. 18. Map of Carboniferous poles and equator (after Howell and Martinez, 1957). Sample locations are indicated by symbols enclosing "L." Used by permission of "Geophysics."

field, is that stability is not so easily established.

Figure 28 is a Schmidt net plot of the data from 14 samples from the Weches (Eocene) of east Texas. It is particularly interesting that the normally magnetized samples, indicated by open circles, are from either very oxidized glauconitic sandstone or iron ore derived from weathering of the glauconite. The three inversely magnetized samples indicated by solid circles are from relatively unoxidized glauconitic sandstone. This would seem to suggest a reversed field during this part of Weches time.

A good example quite often found of scattered magnetization of sedimentary rocks is shown by the plot in figure 29 of a large number of samples obtained from the Oakville formation of Miocene age from

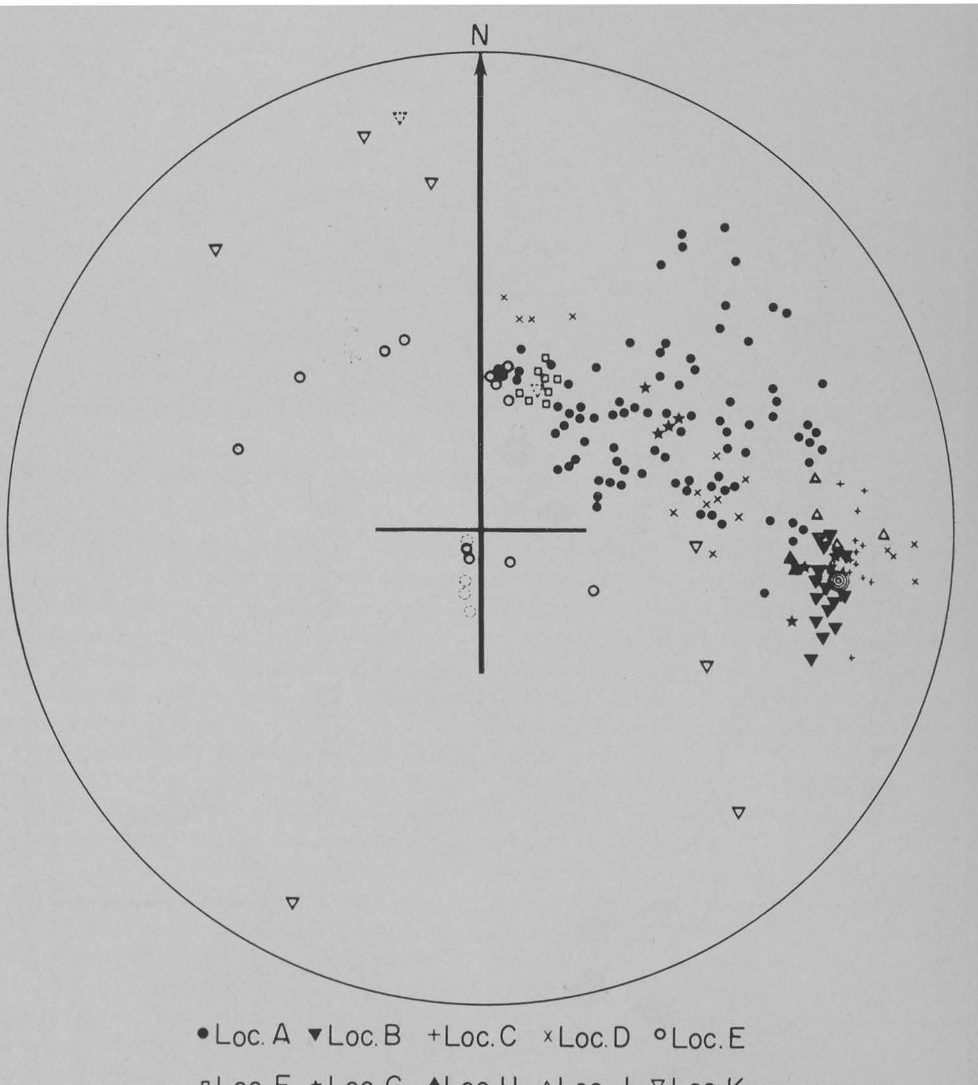


FIG. 19. Point Peak (Cambrian) magnetic vectors, plotted on Schmidt net, all points being on lower hemisphere with exception of dotted points on upper hemisphere (after Howell and Martinez, 1957). Used by permission of "Geophysics."

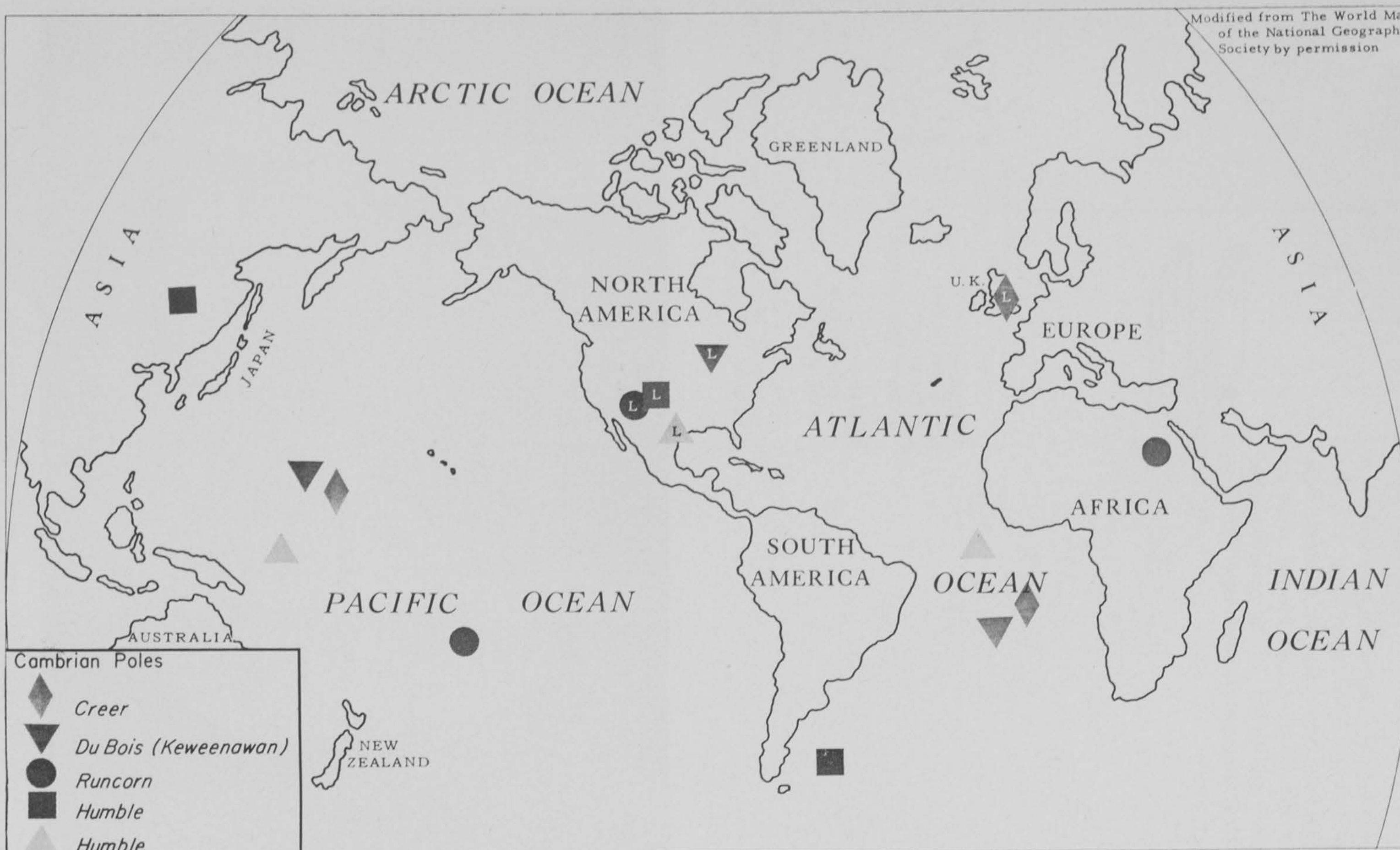


FIG. 20. Map of Cambrian pole locations (after Howell and Martinez, 1957). Used by permission of "Geophysics."

the Texas Gulf Coast. Even here there is a suggestion of some agreement with the present field. Since all of the published data place the Miocene and present poles in similar positions, it is indeterminate whether this is an initial magnetization or not.

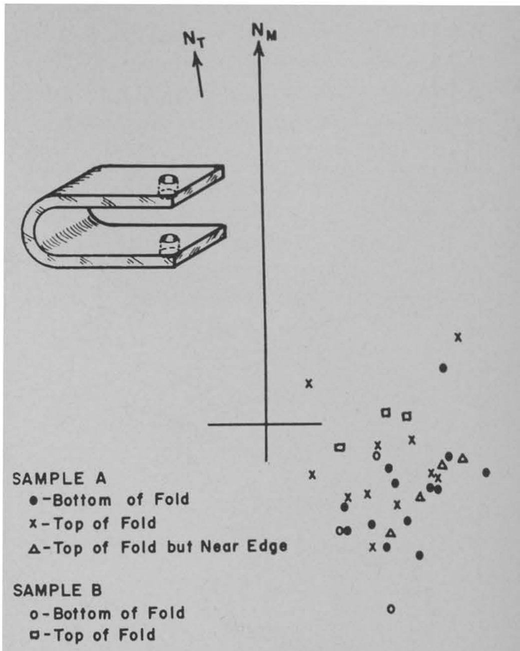


FIG. 21. Lower hemisphere plot of magnetic vectors of penecontemporaneous folds in the Smithwick formation.

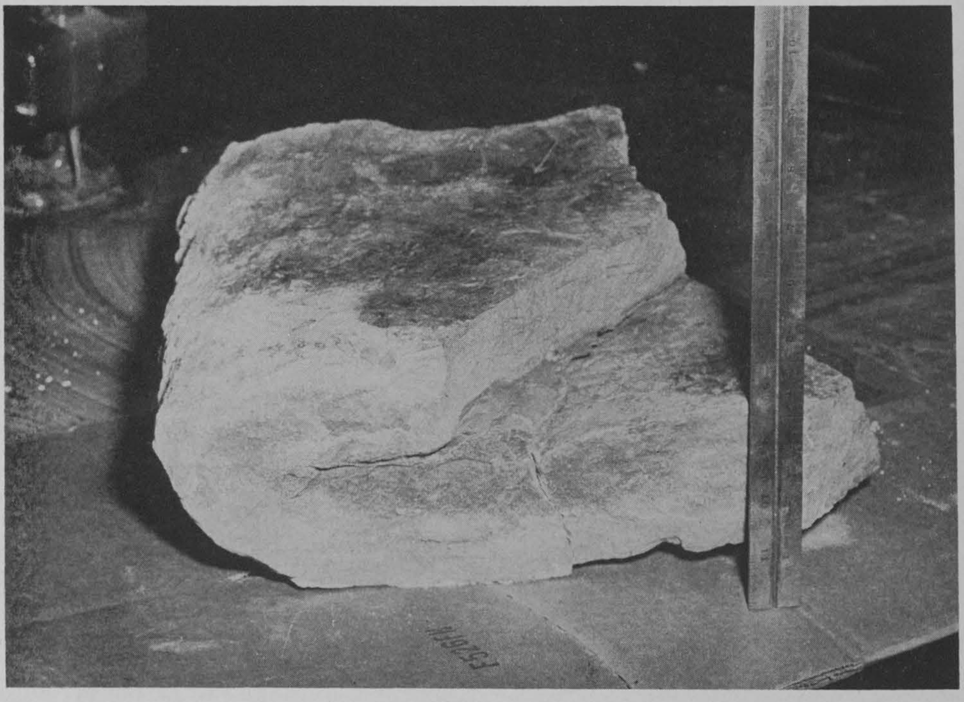

FIG. 22. Sample from a penecontemporaneous fold in the Smithwick formation.

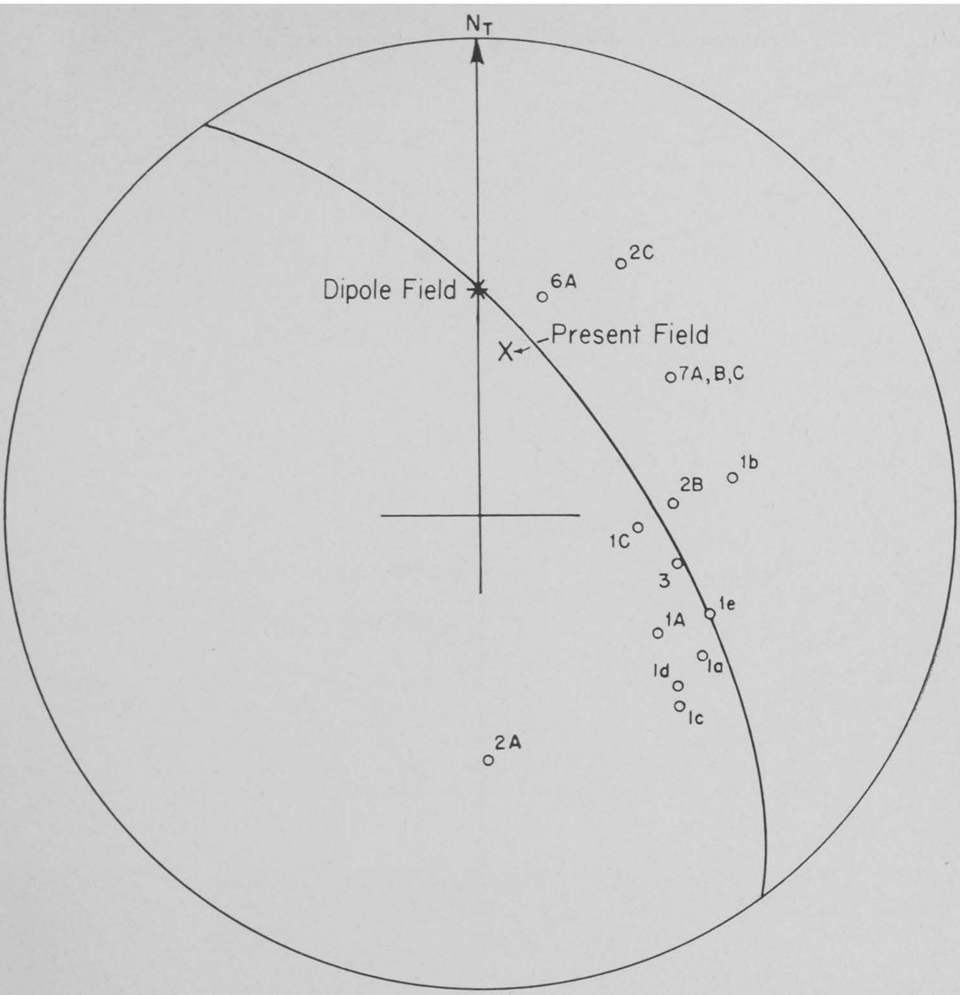


FIG. 23. Lower hemisphere plot of magnetic vectors for fourteen samples of the Smithwick formation.

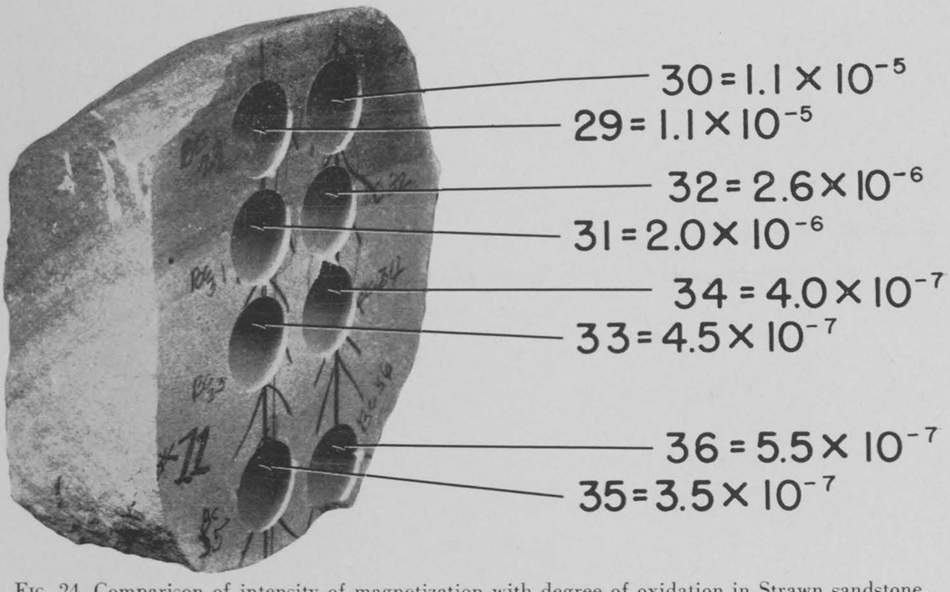


FIG. 24. Comparison of intensity of magnetization with degree of oxidation in Strawn sandstone.

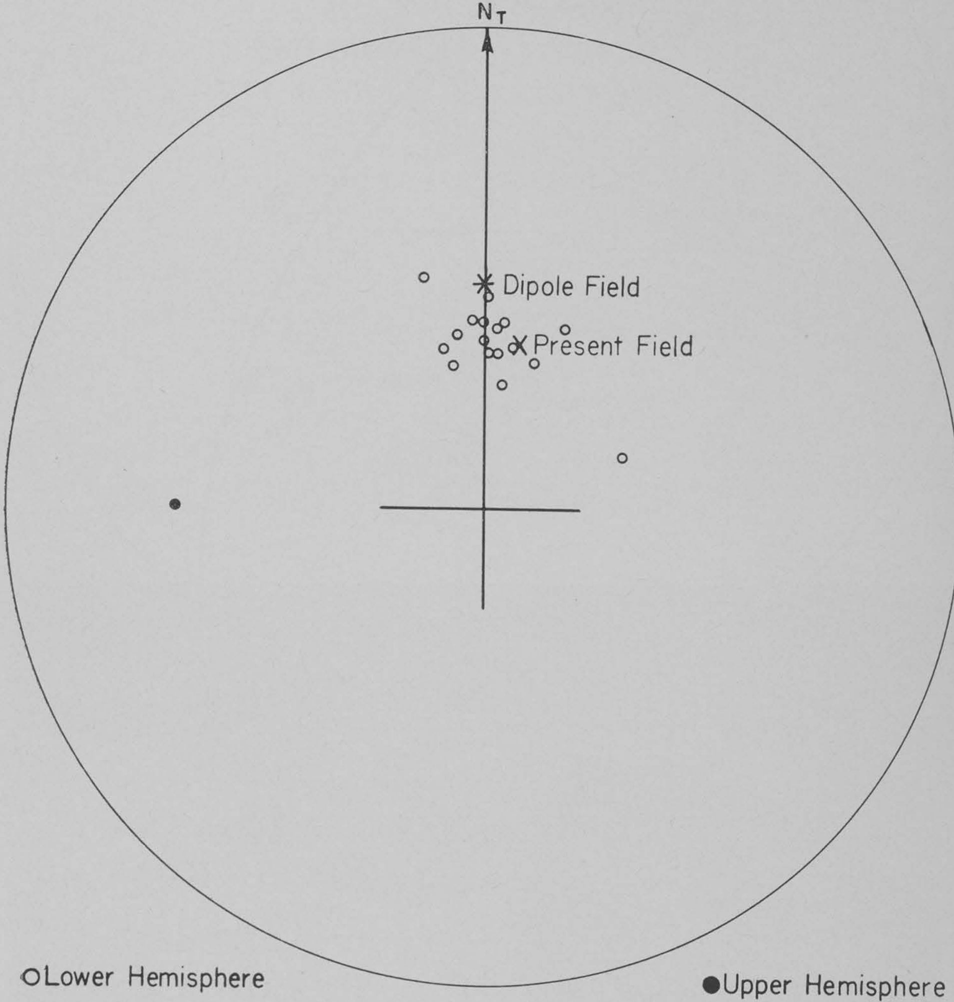


FIG. 25. Plot of magnetic vectors for eighteen samples of the Bell Canyon formation.

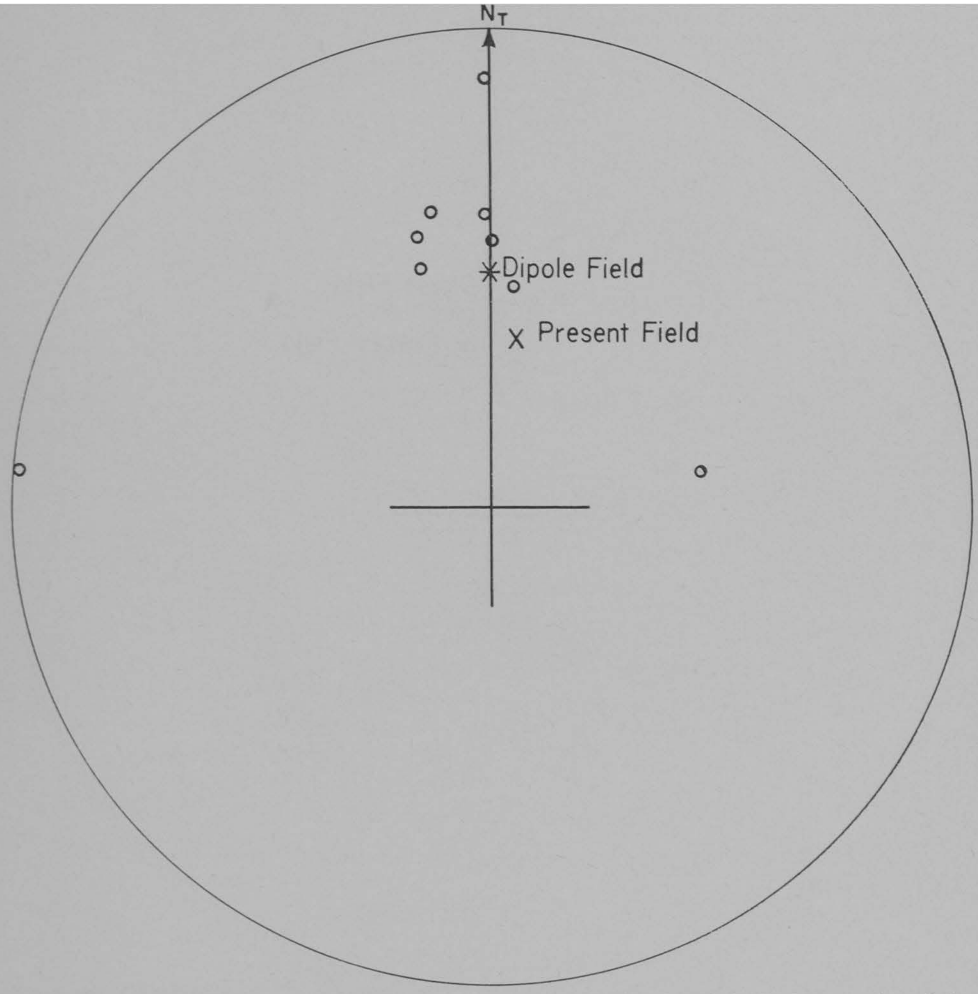


FIG. 26. Lower hemisphere plot of magnetic vectors for nine samples of the Bedias member of the Wellborn formation.

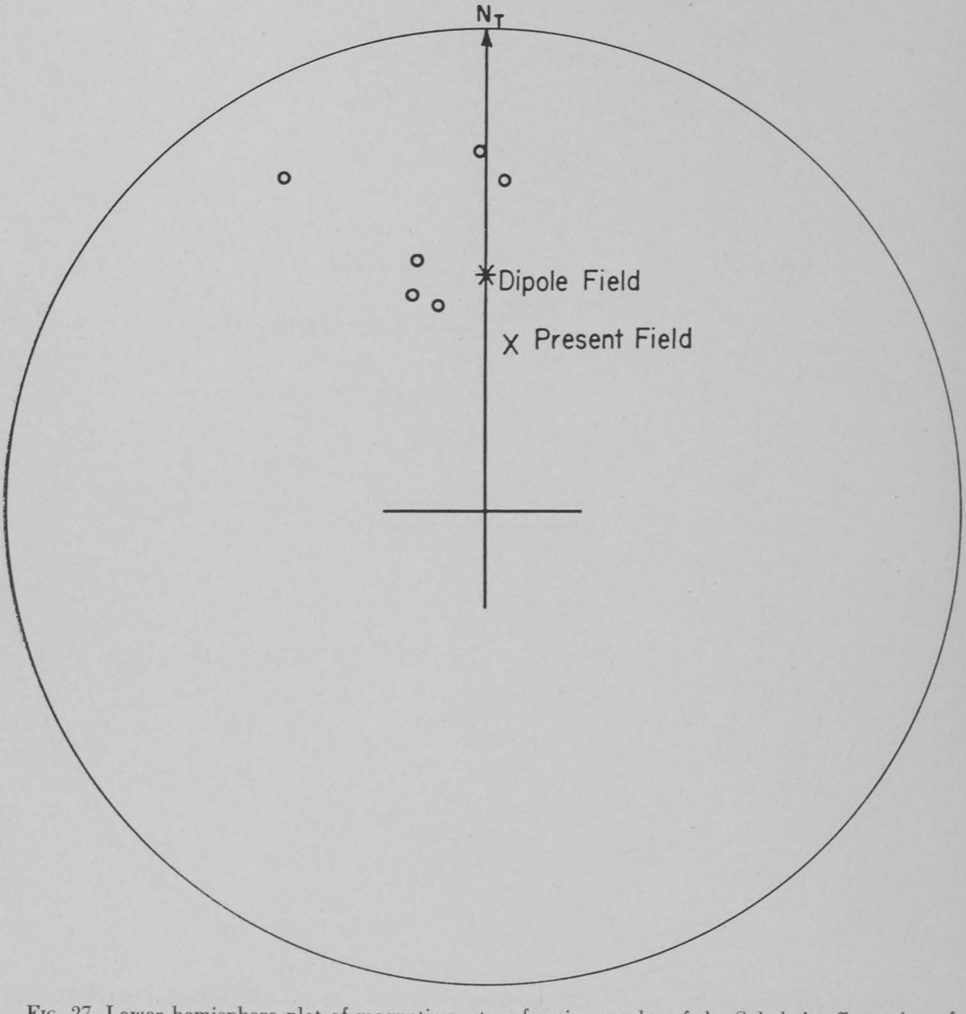


Fig. 27. Lower hemisphere plot of magnetic vectors for six samples of the Soledad tuff member of the Catahoula formation.

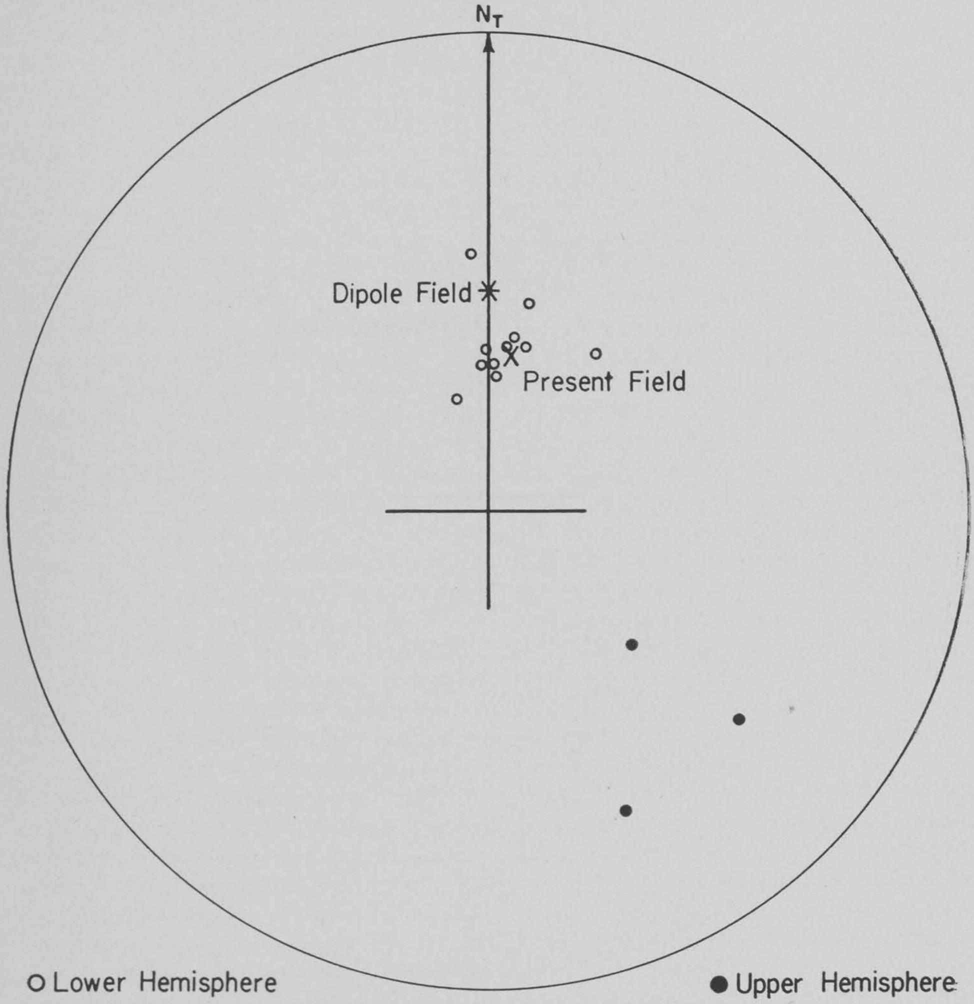


FIG. 28. Plot of magnetic vectors for fourteen samples of the Weches formation.

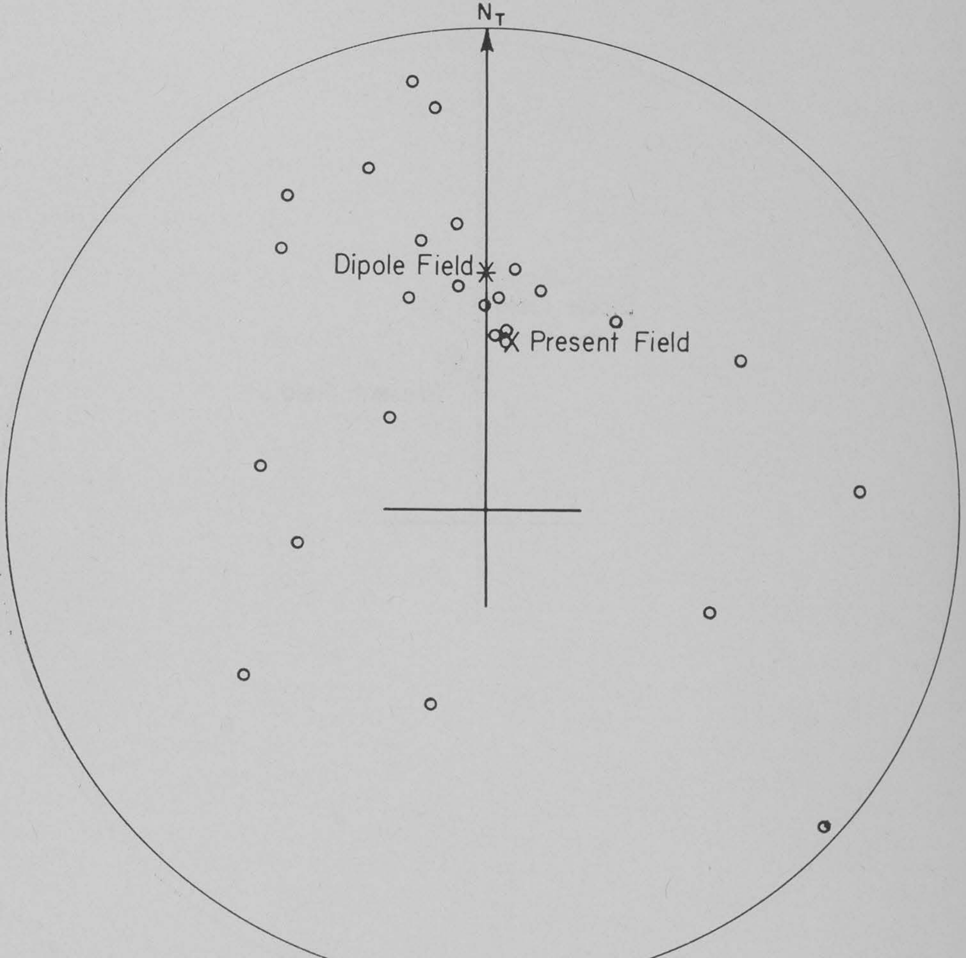


FIG. 29. Lower hemisphere plot of magnetic vectors for twenty-six samples of the Oakville formation.

# PALEOMAGNETIC STUDY OF SOME TERTIARY VOLCANICS IN THE BIG BEND NATIONAL PARK

The next discussion is a brief summary of the results of a paleomagnetic study of some of the Tertiary volcanics in the Big Bend National Park. The principal object of this study was to test the use of such data as a correlation tool as suggested by Runcorn (1956b) and others. Drs. John T. Lonsdale and Ross A. Maxwell. of the Bureau of Economic Geology at The University of Texas, were responsible for relating the sampling program to the geology of the area. Most of the field work was conducted jointly with Dr. Maxwell and to some extent with Dr. Lonsdale. They have been engaged for some time in mapping this area, but their work is as yet largely unpublished. Without their guidance, this study would not have been significant. The volcanic rocks studied are included in the Chisos volcanic series of upper Eocene and Oligocene age. This sequence has been described by Lonsdale et al. (1955) as follows: "The part of the Tertiary sequence in the park area which contains abundant pyroclastic rocks and lavas has been called the Chisos volcanic series." "Sandstone, conglomerate, and fresh-water limestone also are present but various types of pyroclastic and extrusive rocks are characteristic of the series. A complete uninterrupted section has not been found and the thickness and lithology of the sequence vary greatly from place to place within the series." Extensive faulting and complex stratigraphic changes laterally make correlation of the extrusive rocks difficult in many cases from one exposure to another. Lonsdale and Maxwell (verbal communication) have tentatively recognized five basalts, one trachyandesite flow, a flow breccia (including some sediments), and a riebeckite rhyolite flow, in that order from bottom to top. In most cases these units are separated from each other by various thicknesses of Chisos sediments. In many cases, relatively positive correlation of these units can be made from one exposure to another. In some cases, this is difficult to do. Since the basalts are lithologically very similar, they are particularly troublesome. The trachyandesite, the flow breccia, and the riebeckite rhyolite can be distinguished on the basis of lithology, although it is possible that the trachyandesite is a multiflow unit.

For the purpose of this study, samples were obtained from the locations indicated on the topographic map in figure 30. The unit sampled at each location is also indicated. In some instances, the other evidence used for identifying the particular unit may be questionable. These questions are discussed in the following detailed discussion of the data. Figure 31 is a Schmidt net plot of the data from eight samples of what was considered to be the lowest basalt flow (no. 1). In this and the following plots, solid circles represent upward-directed vectors. These samples are from three separate locations. With the exception of one location, the magnetization is in general agreement with the present dipole field. The divergent data show a scatter in results from two samples from a single outcrop. The difference might possibly be explained by erroneous correlation of the basalt at this location with basalt no. 1. Figure 32 is a plot of the data from 21 samples of basalt no. 2, the next highest basalt. These samples are very strongly magnetized but vary greatly in direction from location to location, and in most cases from sample to sample. Location 3 differs in being fairly consistently magnetized in an inverse direction. With this exception, the scatter seems characteristic of this flow. Figure 33 is a plot of the data from basalt no. 3. This flow is characterized by an inverse magnetization. There are certain samples here also that do not fit the general pattern for this flow. Again, the question is raised concerning the accuracy of the correlation of samples from which the data in disagreement were obtained. Of

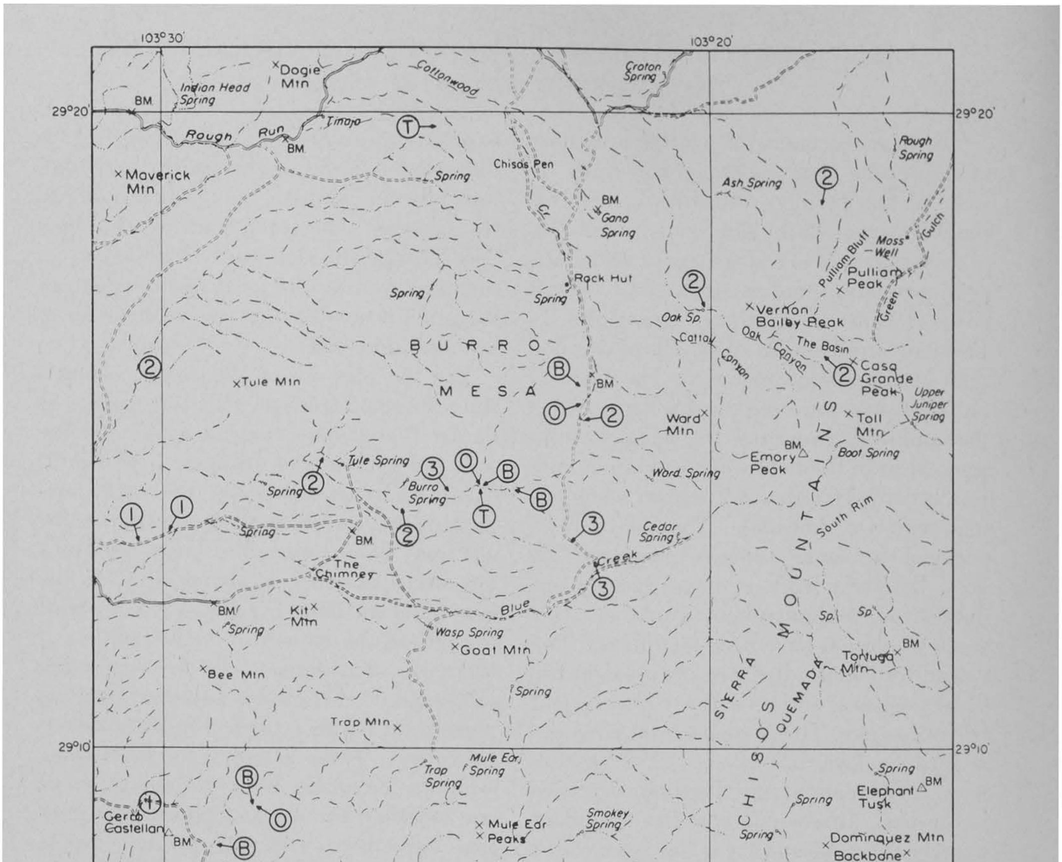


FIG. 30. Sample locations, Big Bend National Park. B = riebeckite rhyolite; O = flow breccia (including some sediments); T = trachyandesite. 1, 2, 3, 4, 5 = basalt flow nos. 1 to 5.

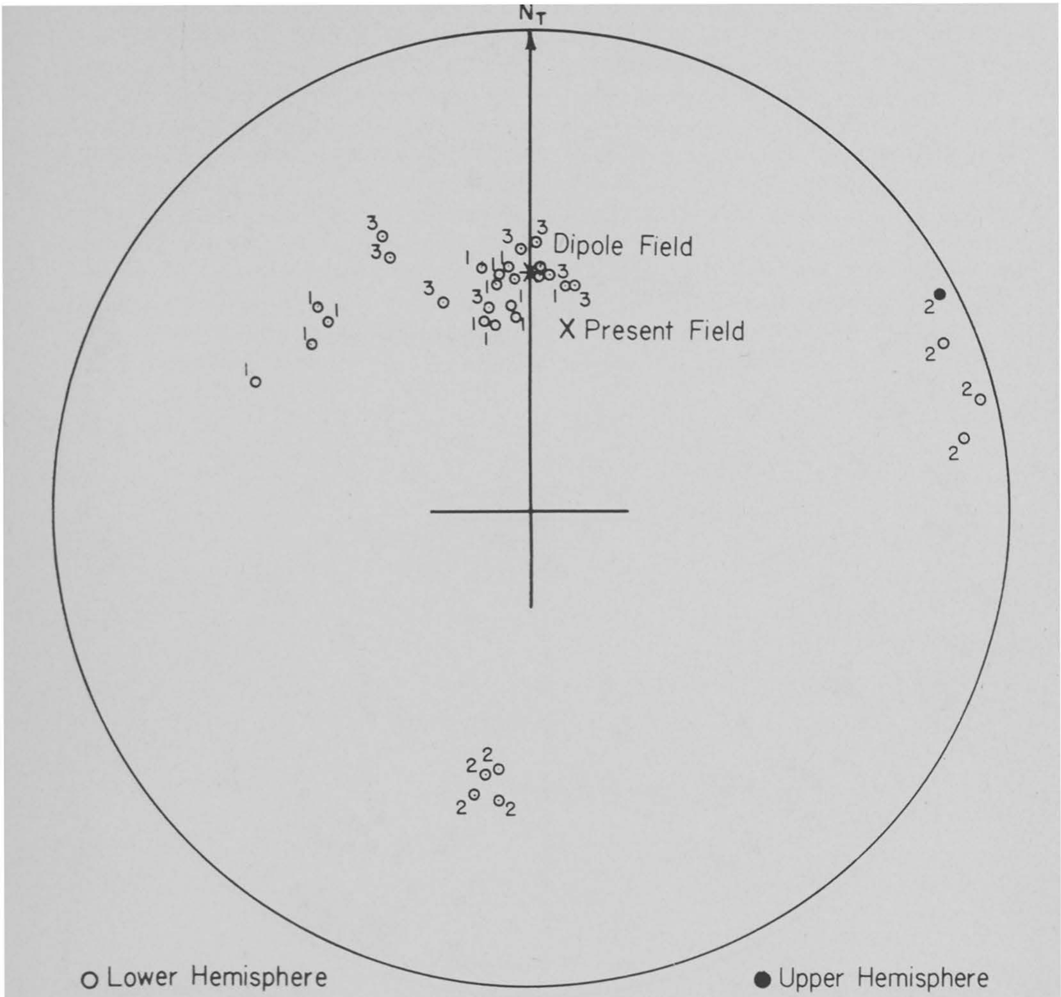


FIG. 31. Plot of magnetic vectors for basalt flow no. 1.

course, unstable samples might also produce this sort of data. The data for basalt no. 4 are shown in figure 34. The assignment of the basalt at location 2 to this flow was stated by Maxwell (veroal communication) to be definitely questionable. However, no very clear picture results, even if these data are excluded. Figure 35 shows that basalt no. 5 is magnetized in a direction near but somewhat different than the present field. However, only one location was sampled. Figure 36 shows the data from 14 samples from six locations of the Tule Mountain trachyandesite. Some of the vectors are scattered but some "home" on the present field. Three possibilities are suggested. First, the Tule may consist of more than one flow and these may have different magnetic characteristics. Second, there may be some penecontemporaneous effects, such as a differential movement of consolidated portions of the flow at temperatures below the Curie point, that resulted in scattered directions of magnetization in this unit. The third possibility is that this rock is somewhat unstable magnetically and that the magnetization directions of many of the samples have drifted back

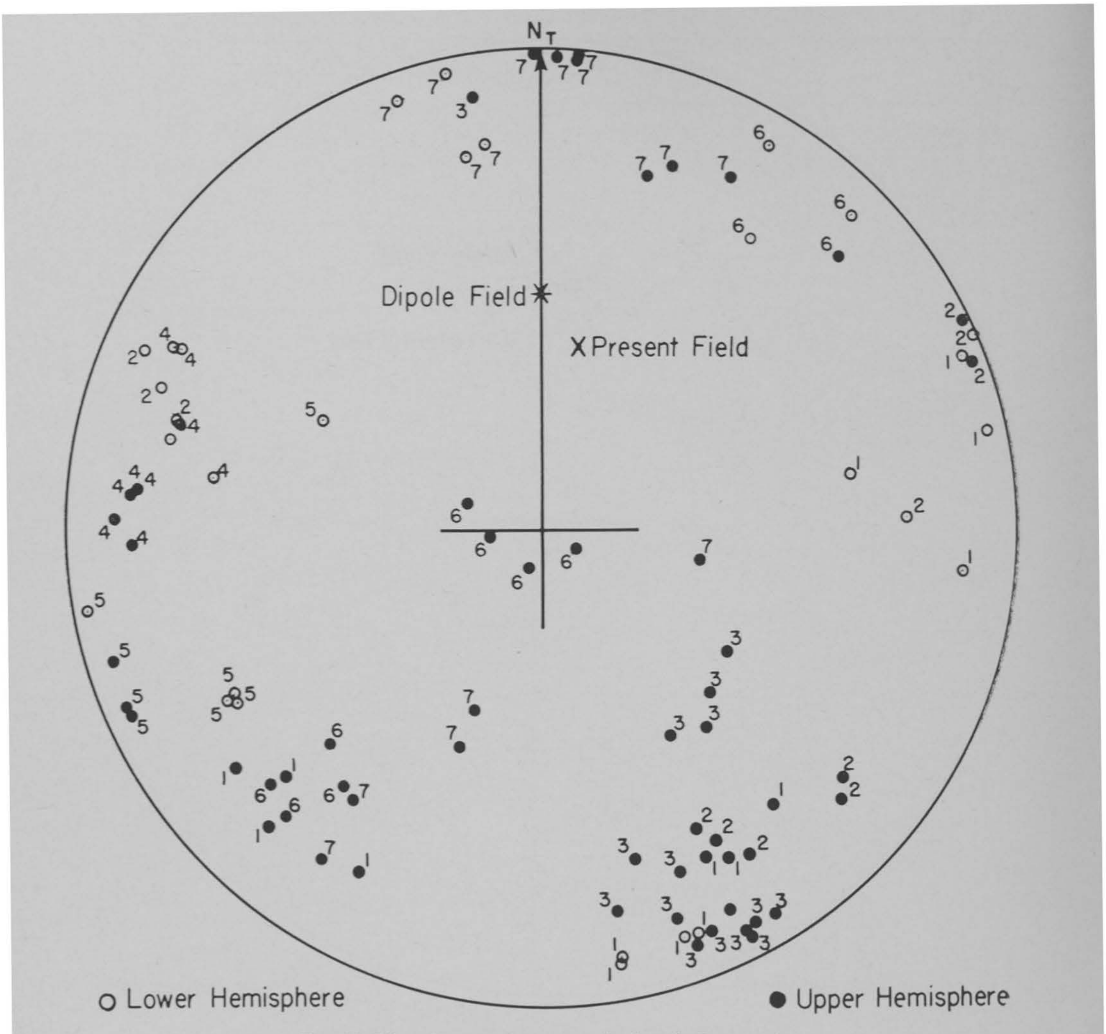


FIG. 32. Plot of magnetic vectors for basalt flow no. 2.

to the direction of the present field, and some have been moving in this direction but have still not reached this position. The fact that a large number of points cluster in the area of the present field is evidence in this regard. It is further suggested that the random location of the other points is evidence that this unit may have initially been magnetized in a direction reverse to the present field, since only from this initial position could the vectors follow an infinite number of great circle paths to the present position of the field. It would seem that a movement from any other point would result in some clustering of the vectors along some single great circle path. The good agreement of data from the same samples is contradictory evidence. This explanation might be applied to some of the other incoherent data. The Burro Mesa riebeckite rhyolite is quite consistently magnetized, and an orange-colored flow breccia unit immediately under the rhyolite considered by Lonsdale and Maxwell (verbal communication) to be a part of the same eruptive cycle seems fairly consistently magnetized, as may be seen on figures 37 and 38. Figure 38 is a plot of the data from 13 samples from five locations of the Burro Mesa riebeckite rhyolite. Figure 37 is a plot of the data from the orange-colored flow breccia unit. With reservations, this sequence of plots lends encouragement to the thesis that remanent magnetism measurements may be used as an aid in correlating volcanic sequences in conjunction with conventional petrographic and field methods. In this particular work, the testing and use of the data have gone hand in hand. Some of the questions that have arisen might have been avoided by working in a less complicated area, but then the use of this tool in field mapping here would have been delayed. Certainly, we must emphasize that for a statistical study of this nature we lack quantities of data. Also, some of the beds show random magnetization. This state may or may not be helpful depending on its frequency of occurrence in the sequence. In the case of basalt no. 2, it appeared to be helpful. On the other hand, there is a good possibility that key beds can be followed, which are consistently magnetized.

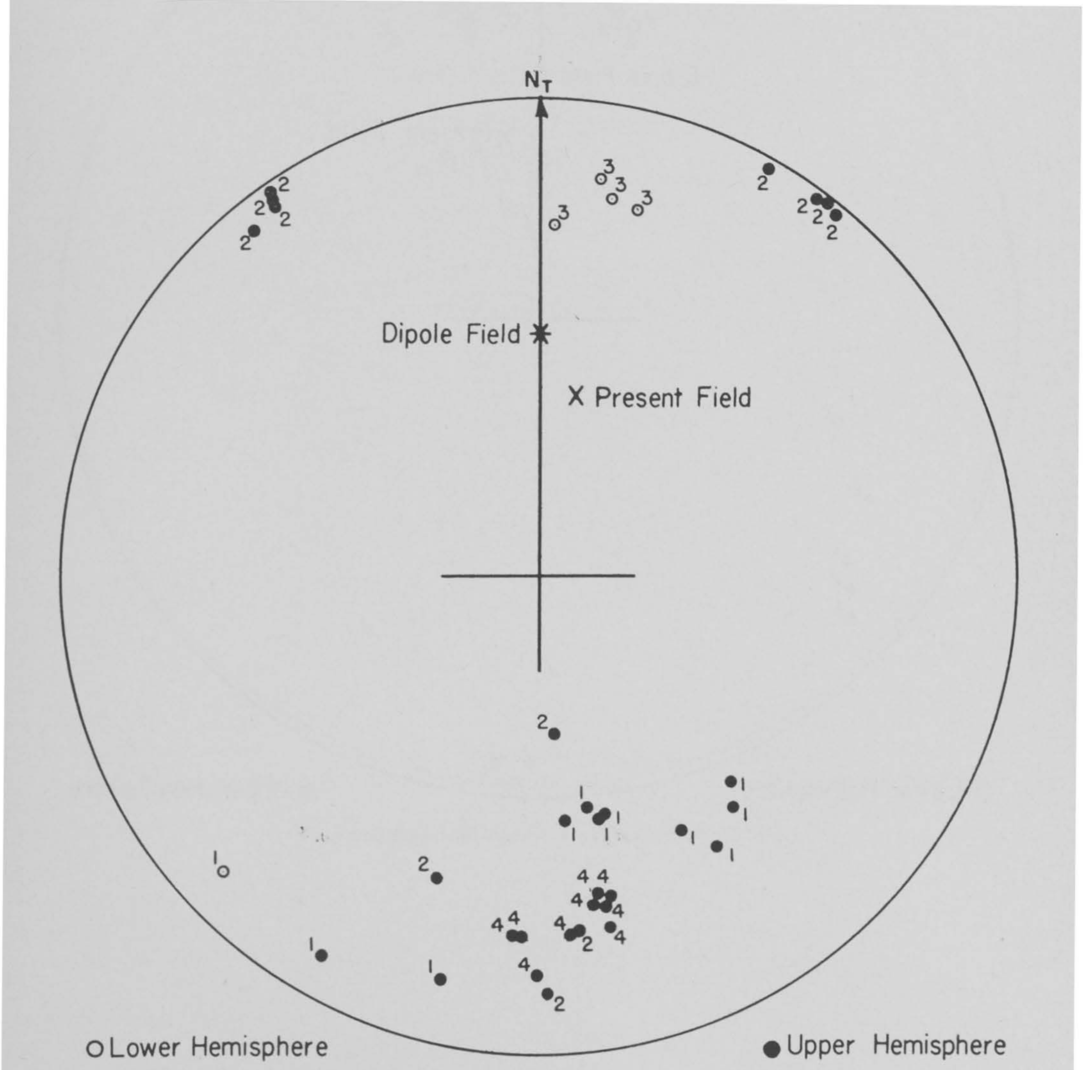


FIG. 33. Plot of magnetic vectors for basalt flow no. 3.

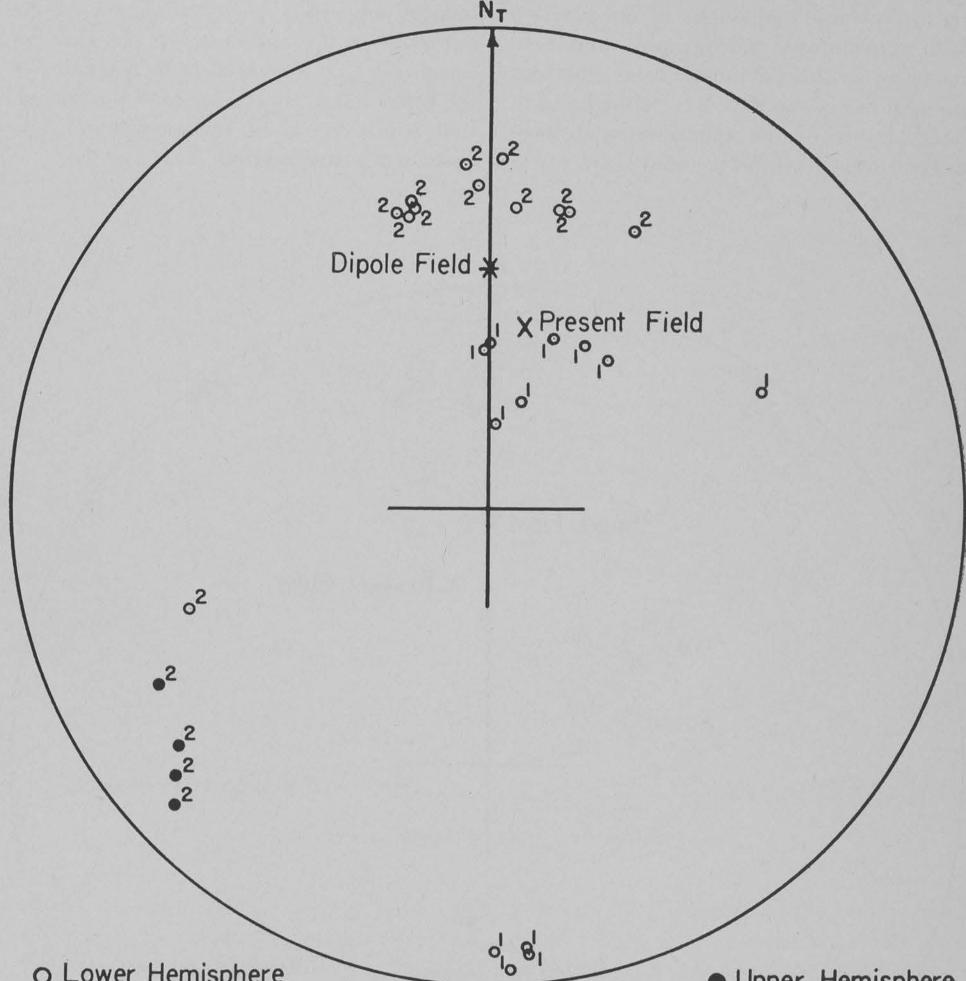


FIG. 34. Plot of magnetic vectors for basalt flow no. 4.

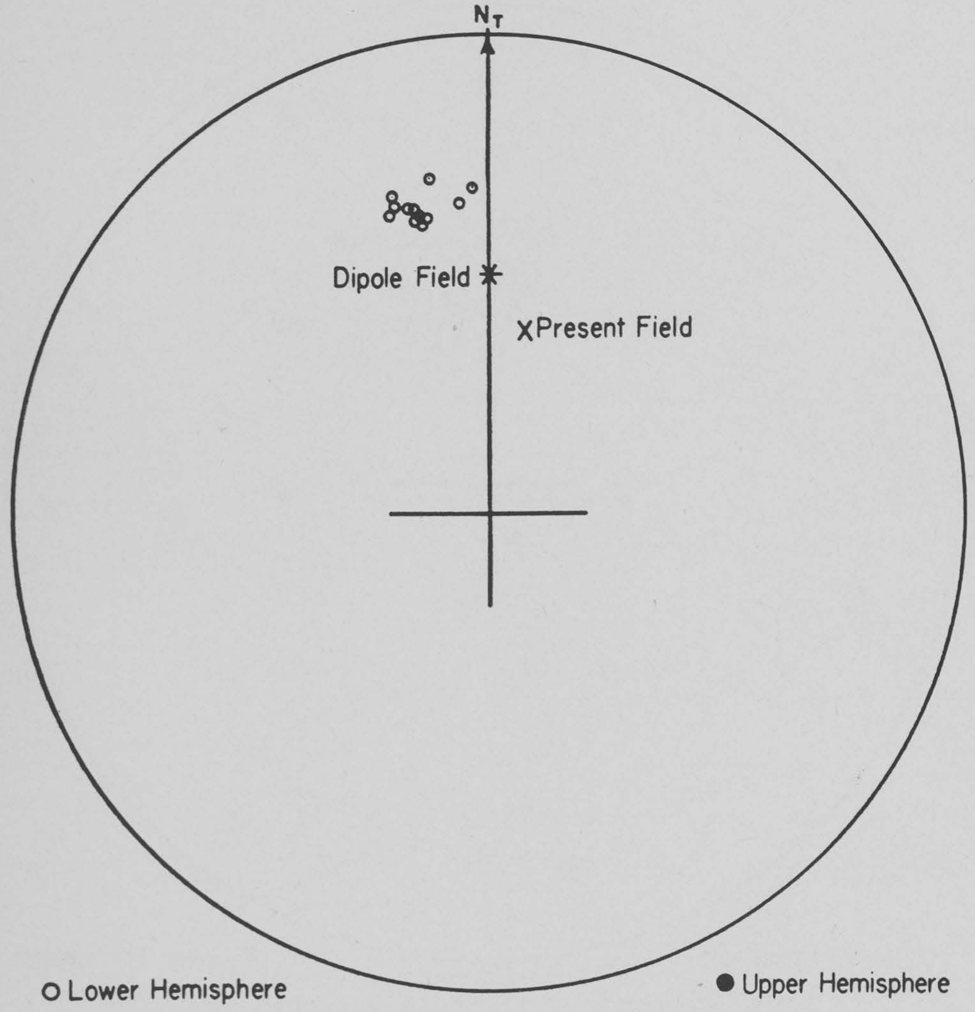


FIG. 35. Plot of magnetic vectors for basalt flow no. 5.

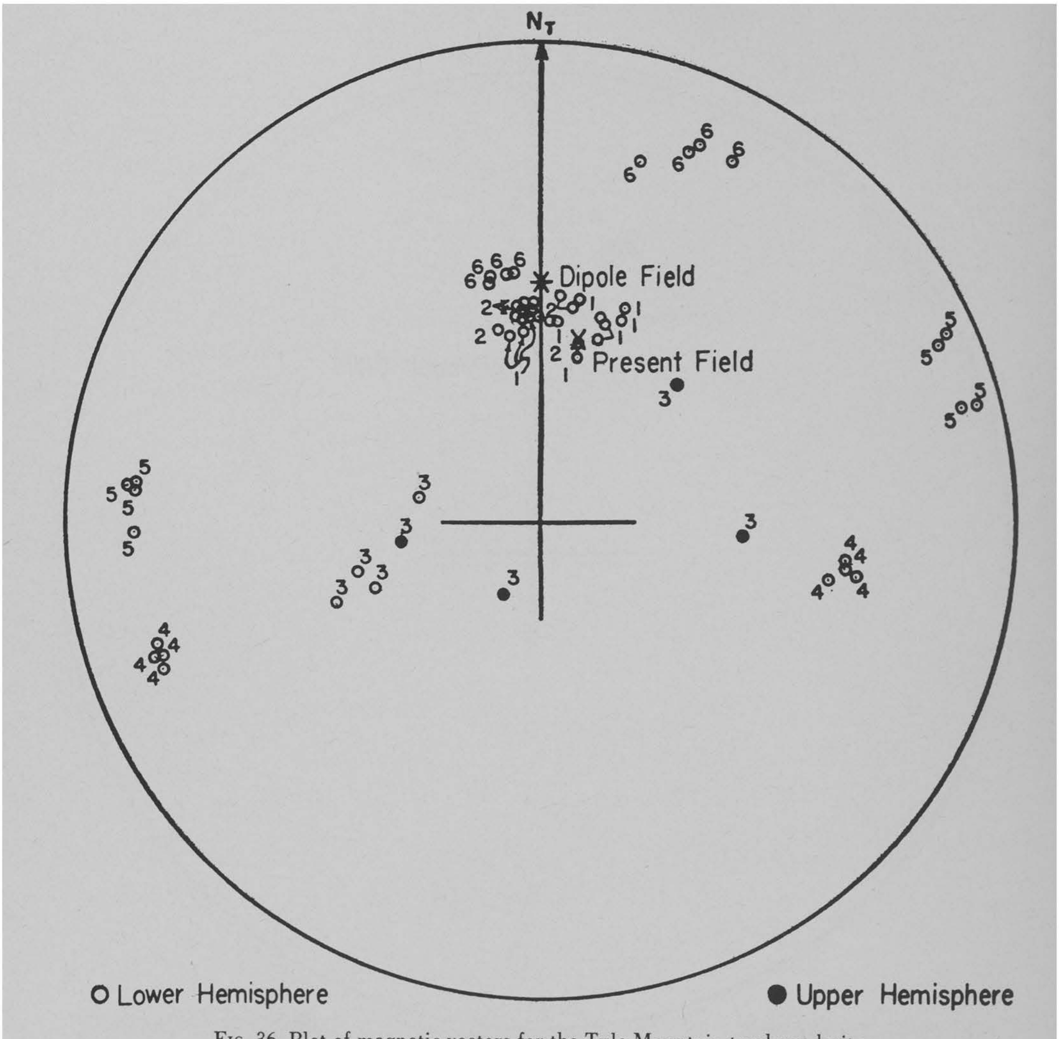


FIG. 36. Plot of magnetic vectors for the Tule Mountain trachyandesite.

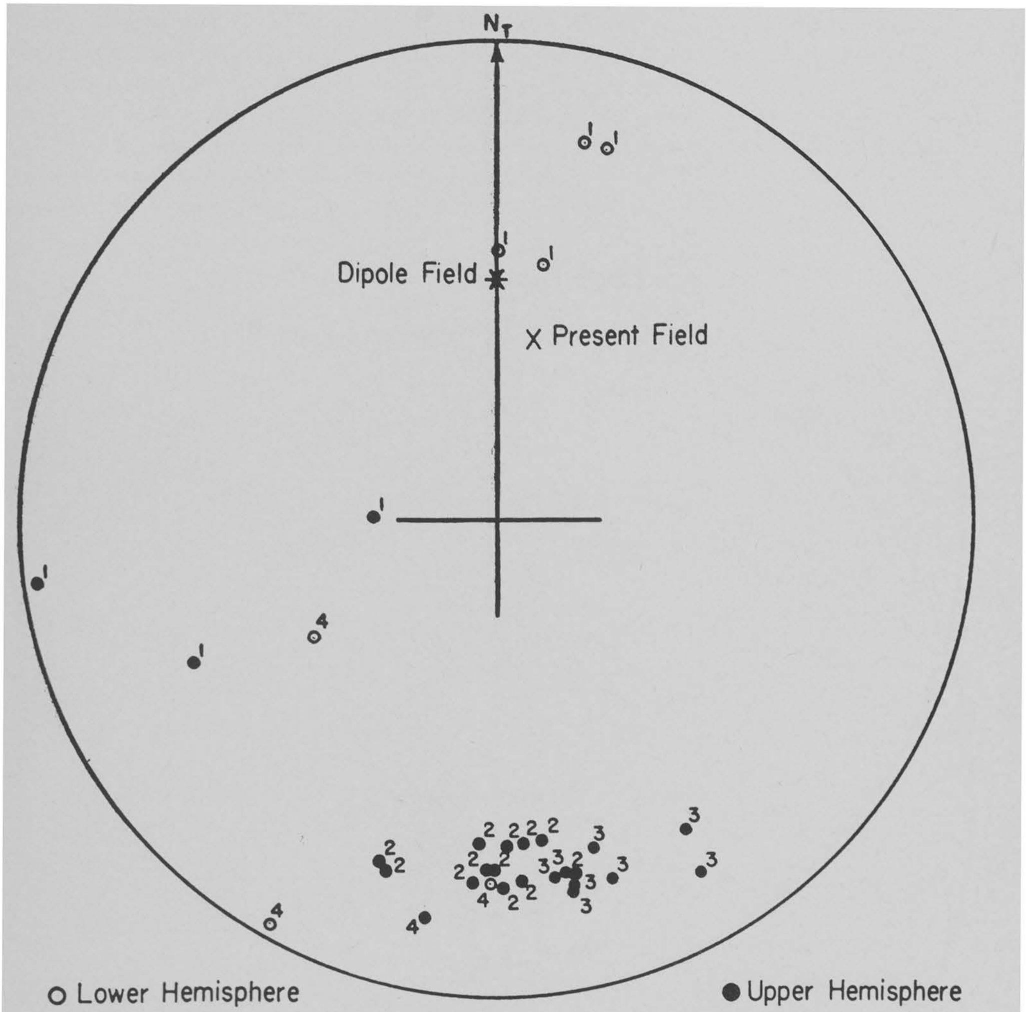


FIG. 37. Plot of magnetic vectors for the orange-colored flow breccia unit.

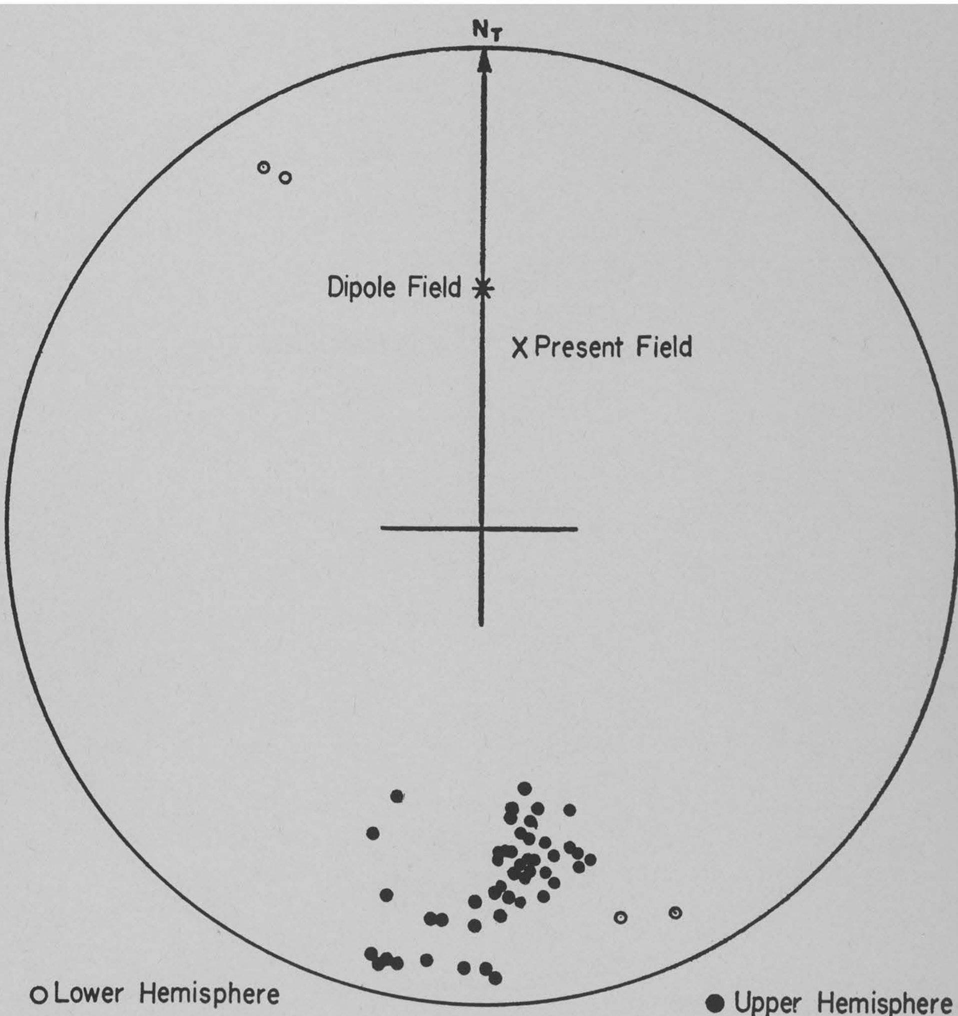


Fig. 38. Plot of magnetic vectors for the Burro Mesa riebeckite rhyolite.

Howell, Martinez, and Statham (1958) have already discussed in some detail results of studies of some metamorphic rocks. The writers would only like to state here that there appears to be a relationship between the direction of magnetization and planar elements in such rocks. (This was first observed by Mr. P. H. Masson of the Humble Oil & Refining Company.) Figure 39 is a plot of magnetic vectors measured in two samples of the Packsaddle schist and shows their relationship to the plane of schistosity. Figure 40 is a plot of magnetic vectors measured in a sample of Valley Spring gneiss and shows their relationship to the plane of foliation. All samples were from outcrops of these two units in the Llano uplift of central Texas.

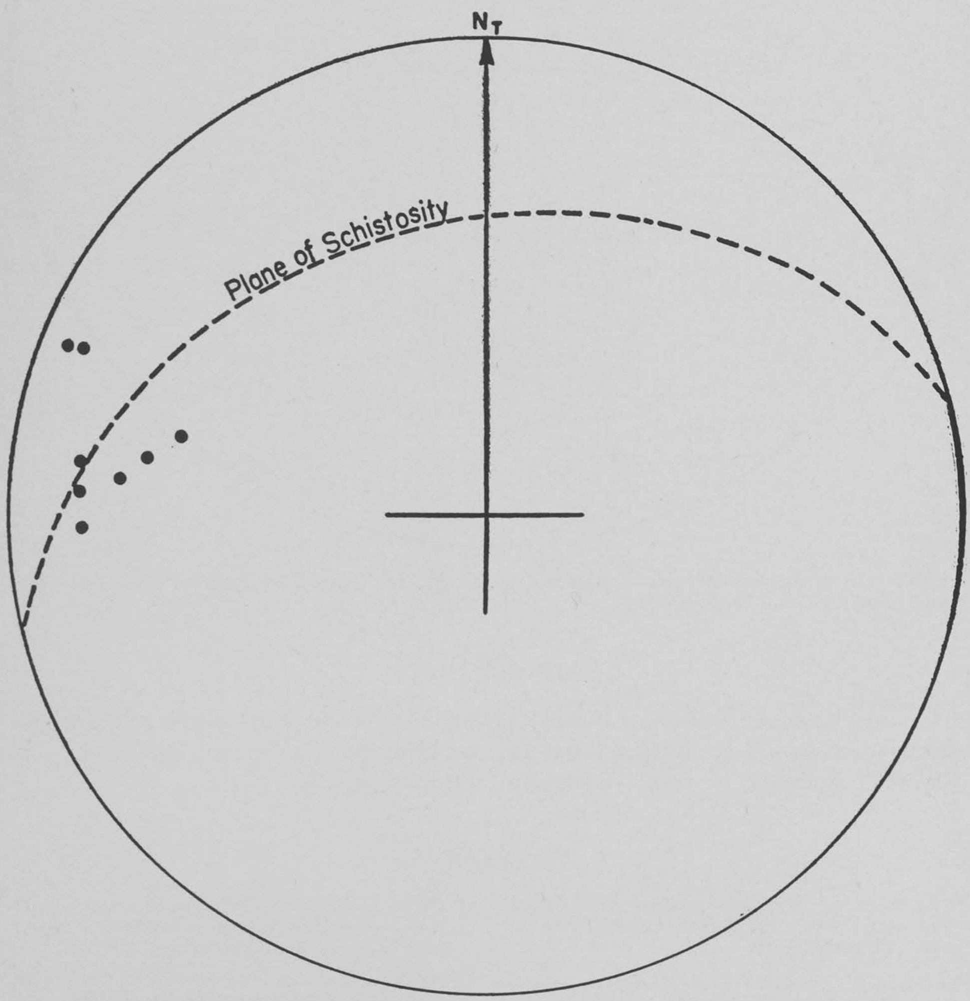


FIG. 39. Upper hemisphere plot of magnetic vectors for the Packsaddle schist. (Modified from Howell, Martinez, and Statham, 1958.)

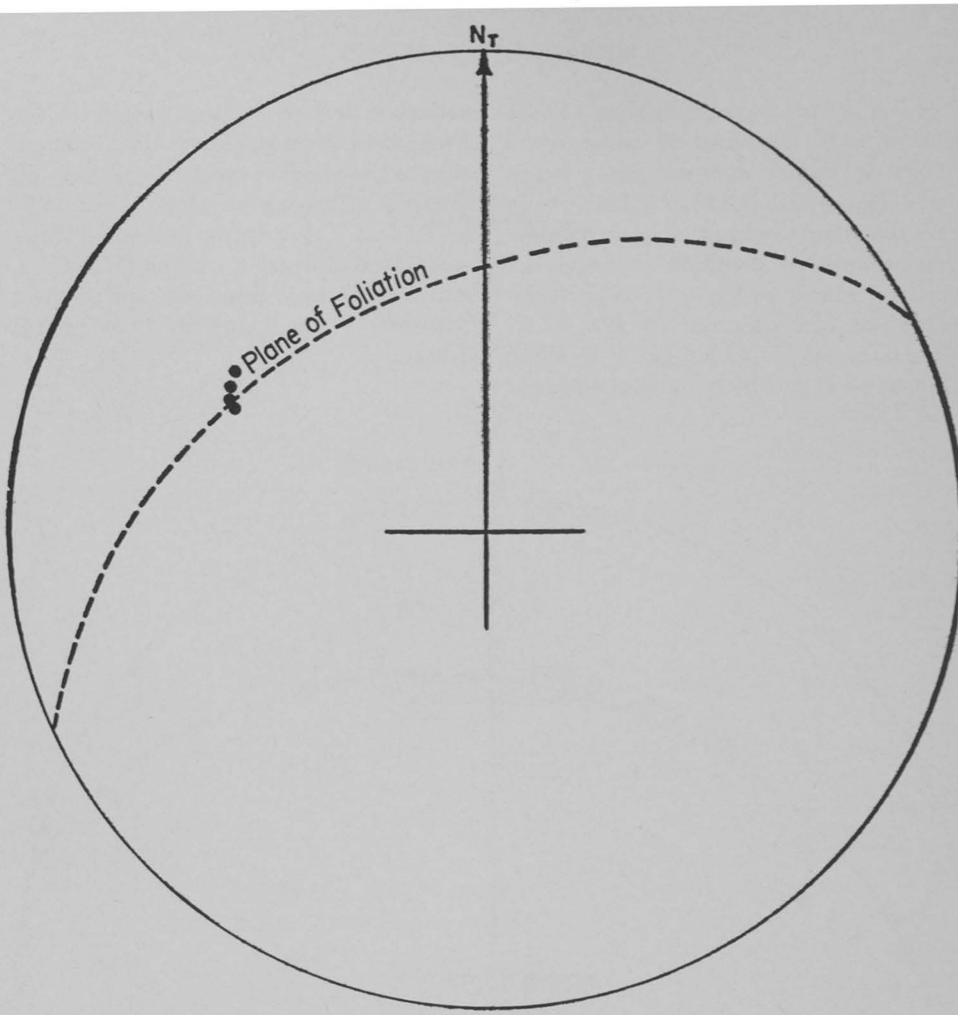


FIG. 40. Upper hemisphere plot of magnetic vectors for the Valley Spring gneiss. (Modified from Howell, Martinez, and Statham, 1958.)

## CONCLUSION

In conclusion, we feel that studies of rock magnetism will be of great value in the broad problems of polar wandering and continental drift, as well as in specific problems such as the correlation of volcanic sequences.

## REFERENCES

BULLARD, E. C. (1949) The magnetic field within the earth: Proc. Royal Soc. London, Ser. A, vol. 197, pp. 433-453.

ELSASSER, W. M. (1945-1947) Induction effects in terrestrial magnetism: Phys. Rev., vol. 69, pp. 106-116 (1945); vol. 70, pp. 202-212 (1946); vol. 72, pp. 821-833 (1947).

GRAHAM, J. W. (1949) The stability and significance of magnetism in sedimentary rocks: Jour. Geophys. Res., vol. 54, pp. 131-167.

HANDBUCH DER PHYSIK (1956) vol. 47, Geophysik I, pp. 482-486, Springer-Verlag, Berlin.

HOSPERS, J. (1955) Rock magnetism and polar wandering: Jour. Geol., vol. 63, pp. 59-74.

HOWELL, L. G., and MARTINEZ, J. D. (1957) Polar movement as indicated by rock magnetism: Geophysics, vol. 22, pp. 384–397.

———, ———, and STATHAM, E. H. (1958) Some observations on rock magnetism: Geophysics, vol. 23, pp. 285–298.

LONSDALE, J. T., et al. (1955) Guidebook for 1955 spring field trip to Big Bend National Park, Texas, West Texas Geol. Soc., p. 47.

MARTINEZ, J. D., and HOWELL, L. G. (1956) Paleomagnetism of chemical sediments: Nature, vol. 178, pp. 204–205.

NAGATA, TAKESI (1952) Reverse thermo-remanent magnetism: Nature, vol. 169, pp. 704–705.

NÉEL, LOUIS (1951) L'inversion de l'aimantation permanente des roches: Ann. Géophys., vol. 7, pp. 90–102.

——— (1955) Some theoretical aspects of rock magnetism: Advances in Physics, vol. 4, pp. 191–243.

RUNCORN, S. K. (1955a) Rock magnetism—geophysical aspects: Advances in Physics, vol. 4, pp. 244–291.

——— (1955b) Paleomagnetism of sediments from the Colorado plateau: Nature, vol. 176, pp. 505–506.

——— (1956a) Paleomagnetic comparisons between Europe and North America: Proc. Geol. Assoc. Canada, vol. 8, pt. 1, pp. 77–85.

——— (1956b) Application of the remanent magnetization of rocks: Geophysical Prospecting, vol. 4, pp. 221–225.

# Deposition and Alteration of the Edwards Limestone, Central Texas[4]

HENRY F. NELSON[5]

ABSTRACT

The Edwards limestone is the uppermost formation in the Fredericksburg group (Early Cretaceous epoch). In the vicinity of the Red River, the group is composed predominantly of terrigenous clastic sediments. To the south, the terrigenous sediments grade into the marls, shell beds, and nodular limestones of the Walnut and Comanche Peak formations. The latter, in turn, grade into the Edwards formation farther south. Near Austin, the Edwards formation constitutes most of the Fredericksburg group.

At various localities in Bell, Coryell, and McLennan counties, the Comanche Peak limestone grades into the Edwards limestone by (1) an increase in grain size, (2) a gradual increase in the number of rudistids in the upper part of the Comanche Peak limestone, (3) transition of massive nodular limestone into well-bedded limestone, and (4) intertonguing of nodular limestone with rudistid limestone.

The Edwards formation is 16 feet thick north of Gatesville in Coryell County. It increases in thickness to the south and east reaching a maximum thickness of 124 feet near Moffat in northern Bell County. South of Moffat, it decreases in thickness. It is 68 feet thick at locality 14-T-8 southwest of Belton. Variations in thickness of the Edwards formation are due primarily to facies changes of the Edwards limestone into the Comanche Peak limestone. However, topographic relief, due either to local reef growth in the Edwards limestone or erosion of the limestone prior to deposition of the overlying formations, probably caused some variation in thickness.

The Edwards formation is unconformably overlain by the Kiamichi and Duck Creek formations. Evidence for an unconformity includes (1) oxidation and case-hardening of the top of the Edwards limestone, (2) occurrence of small pits and bore holes filled with Kiamichi shale in the top of the Edwards limestone, (3) onlap of successively higher lithologic units of the shale upon the Edwards formation, and (4) onlap and pinchout of the shale around rudistid reefs. There is no evidence of gradation between the two formations. The Kiamichi shale pinches out in southeastern Coryell County along a line extending from Whitson toward Gatesville.

In the area of this study, the Edwards formation is a reef complex made up of massive rudistid biohermal and biostromal reefs that grade laterally into well-bedded inter-reef deposits. Biohermal reefs are composed of a mass of rudistids and associated organisms embedded in a very fine-grained matrix. Three faunal zones can be frequently recognized. A coral zone in which *Cladophyllia* is prominent occurs at the base of the reefs. The *Cladophyllia* zone grades upward into a zone of *Toucasia* and *Monopleura*. The *Monopleura-Toucasia* zone grades upward and outward from the reef core into the zone of *Caprinuloidea*, *Eoradiolites*, and *Chondrodonta*. The biohermal reefs range from a minimum thickness of 9 feet to a maximum known thickness of 55 feet. The reef cores grade laterally into more fragmental

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[4] This paper was presented at the meeting in Austin in October 1959 and appears on pages 21-95 of The University of Texas Publication 5905, "Symposium on Edwards Limestone in Central Texas." For this reason, abstract (reprinted) only is included in this present publication.

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flank beds that dip away from the cores with inclinations as great as 35 degrees. In some places, the biohermal reefs apparently stood at least 20 feet above the surrounding sediments.

The inter-reef sediments are composed of well-sorted calcilutites, calcarenites, and poorly sorted shell debris. Most particles are well rounded and are composed mainly of "original" shell fragments, recrystallized shell fragments, and opaque grains. The particles are cemented with clear calcite that is believed to be an original precipitate rather than a product of recrystallization. The chert in the inter-reef facies is a primary deposit.

Primary dolomite occurs as beds and as crystals disseminated in limestones and chert. Dolomite also occurs as a diagenetic mineral in the matrix of limestones, in the body chambers and shell walls of fossils, in bore holes, and in voids in reef limestones.

In Bell and southeastern Coryell counties, south of the pinchout of the Kiamichi shale, the Edwards formation has been altered by post-lithification processes which include solution, recrystallization, cavity filling, dolomitization, and silicification. The resulting limestones are characteristically mottled shades of brown, yellow, and pink. They are hard dense crystalline limestones that occur as beds, concretions, and irregular-shaped masses. Post-lithification dolomite is soft, very finely crystalline, and has excellent intercrystalline porosity, except where it has been cemented by subsequent precipitation of calcite in the pores.

This study and a previous study [Feray, D. E., and Nelson, H. F. (1956) Nature of porosity and permeability in the Edwards formation, Texas (abst.): Amer. Assoc. Petr. Geol., Program of 41st Ann. Meeting, Chicago, pp. 14–15] have shown that post-lithification dolomite occurs where the Kiamichi shale is thin or absent and that dolomitization took place prior to deposition of the Duck Creek limestone. The time when the crystalline and silicified limestones formed has not been positively established. Some of them formed after dolomitization. Extensive chalkification of the Edwards limestone appears to be related to present-day topography.

During the Early Cretaceous epoch, the rudistids and associated organisms formed one of the most extensive reef complexes in geologic history. At the beginning of Fredericksburg time, the fauna began to migrate northwestward from the main reef trend. As they migrated, they transgressed the Fredericksburg group and formed a reef complex along the west side of the Tyler basin. The reef complex, which is described in this study, effectively subdivided the lagoon behind the main reef trend into two parts: the Austin lagoon in which rudistid biostromes, granular limestones, and chert (Edwards) were formed and the Tyler lagoon in which the Paluxy, Walnut, and Comanche Peak formations were deposited. The Fredericksburg age was brought to a close by regional uplift, but before uplift took place, reef growth had ceased and sedimentation had essentially filled the inter-reef basins to the crests of the reefs. Uplift was apparently not very great. Following uplift, the Edwards limestone was subjected to post-lithification alteration that developed new types of carbonate rocks.

# Geology of the Texas Panhandle

JOHN H. NICHOLSON[6]

## ABSTRACT

The Amarillo uplift in the center of the Panhandle and the Matador arch near the southern limit are dominant structures. The Amarillo uplift connects with the Wichita Mountains of Oklahoma and is én echelon with the Bravo dome. an element of the Sierra Grande uplift of New Mexico. The Matador arch parallels the Red River arch of north Texas and Milnesand dome of New Mexico. There are three basins in the Panhandle: deep Anadarko basin north of the Amarillo uplift; shallow Dalhart basin in the northwest Panhandle; and Palo Duro basin between the Amarillo and Matador structures. Secondary structures oblique to the Amarillo uplift occur in adjacent basins.

The Texas Peninsula was a broad arch until Mississippian time; structural activity responsible for basins and uplifts commenced in Late Mississippian and culminated in Middle Pennsylvanian (Des Moines) time.

A relatively complete section of pre-Pennsylvanian rocks occurs in the Anadarko basin; the Palo Duro and Dalhart basins contain only Cambrian, Ordovician. and Mississippian sediments. Early and Middle Pennsylvanian clastics eroded from rising structures were trapped in adjacent basins, with some carbonates deposited away from clastic sources. Late Pennsylvanian carbonate deposition dominated shelf areas of subsiding basins; fine clastics accumulated in the center of basins. Shallow water deposits covered the basins in Late Pennsylvanian followed by subsidence and renewed carbonate deposition during Wolfcamp (Permian) time; post-Wolfcamp Permian deposits are evaporites and terrigenous clastics. Regional uplift was followed later by deposition of Triassic and Cenozoic nonmarine rocks.

## INTRODUCTION

The Panhandle of Texas is situated in the southern part of the Great Plains region (fig. 41). In 1919 natural gas was discovered in what is now the Panhandle field on the crest of the buried Amarillo Mountains, and this discovery led to a development program that uncovered the largest single gas field in the world. Since that time the petroleum industry has located numerous oil and gas fields, and petroleum is now one of the major economic assets of the region.

Subsurface geology in the Texas Panhandle has been discussed in many publications since Charles N. Gould (1907, pp. 14–15, 18–21) recognized the tectonic nature of folds in Upper Permian beds along the Canadian River. Gould, truly a pioneer geologist, introduced the terms "Amarillo Mountains" (1923. p. 552), "Anadarko basin" (1924, p. 324), and "Palo Duro basin" (*in* Gould and Lewis, 1926, p. 14). Papers by many subsequent workers, who have contributed to the understanding of the Paleozoic rocks beneath the flat Tertiary "caprock" and adjacent dissected Triassic and Upper Permian red beds, are listed in Sellards (1933) and Girard (1959). Published papers have touched on many geologic aspects, from oil field studies to paleogeography and paleontology. Rogatz's (1935, 1939) comprehensive papers on the geology of the large Panhandle oil and gas field outlined the main features of one of the most intensively drilled structures in the world. Roth (1955) and Totten (1956) discussed

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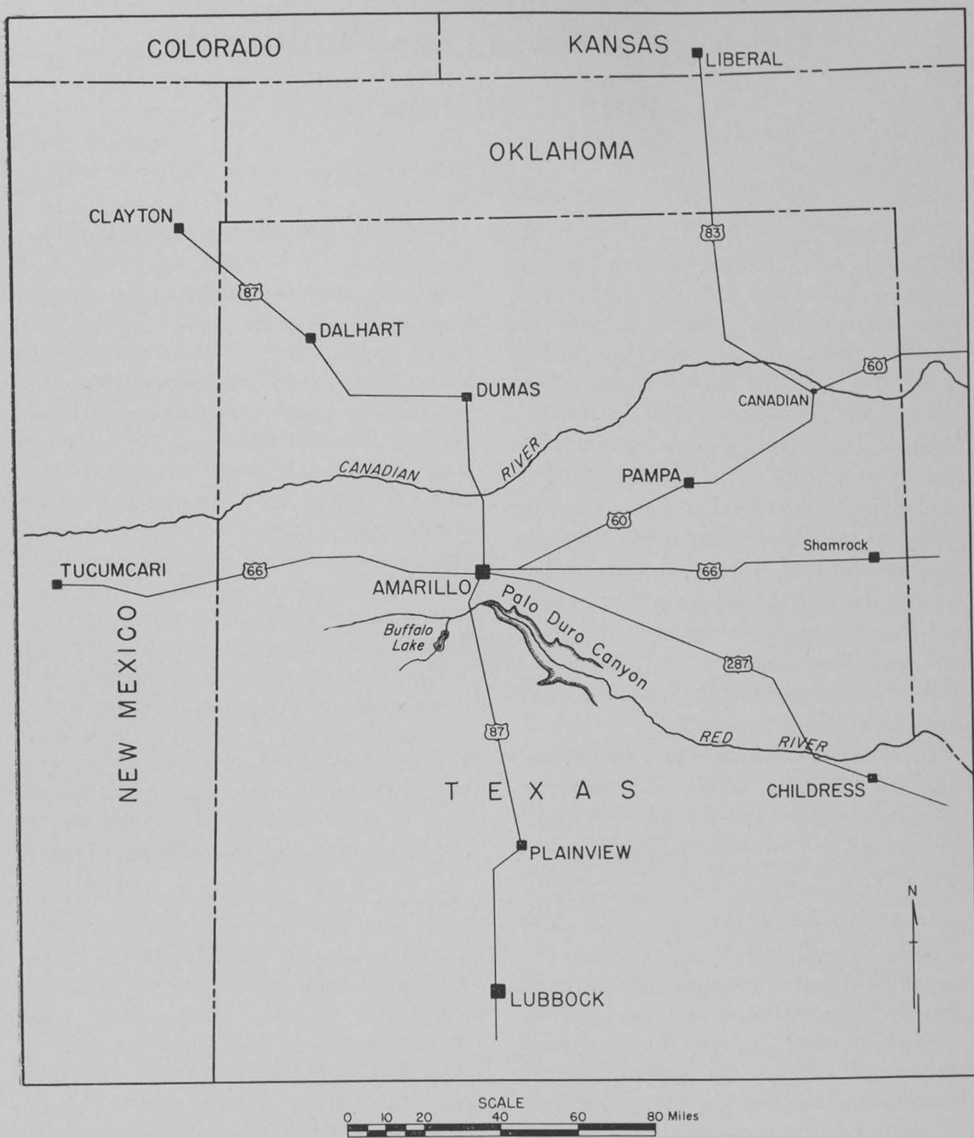


FIG. 41. Texas Panhandle and adjacent areas.

the general stratigraphy, structure, and geologic history of the province. Totten's paper (pp. 1965–1967) also contains a useful bibliography.

In order to make a geologic study of the Panhandle region, most emphasis must be placed on subsurface data derived from wells in the area. Without this information, geological knowledge in the area would indeed be superficial. In some parts of the Panhandle, wells are closely spaced and correlations are very reliable, but unfortunately in other parts spacing is distant and correlations are doubtful.

## STRUCTURE

The basement structure of the Panhandle is dominated by two main trends of folding (fig. 42). The Amarillo Mountain uplift, the more prominent of the two, trends northwesterly across the center of the Panhandle, and the Matador arch or uplift trends in a westerly direction across the southern limit. Complementing these two features are the Wichita Mountains of southwest Oklahoma, in trend with the Amarillo Mountains, and the Oldham nose, more commonly called the Bravo dome, which plunges to the southeast and is also én echelon with the Amarillo Mountains. On trend with the Matador arch are the Red River arch of north-central Texas and the Milnesand dome of New Mexico and Texas. Dominating the area immediately west of the Texas Panhandle is the massive Sierra Grande uplift of northeast New Mexico.

In conjunction with the two major systems of folds is a less prominent group striking southward at an angle from the Amarillo Mountains into the Palo Duro basin, and another group striking northwestward across the Anadarko basin. The Anadarko basin is extremely deep adjacent to the mountain belt, due to the prominent fault system on the north flank of the Ama-

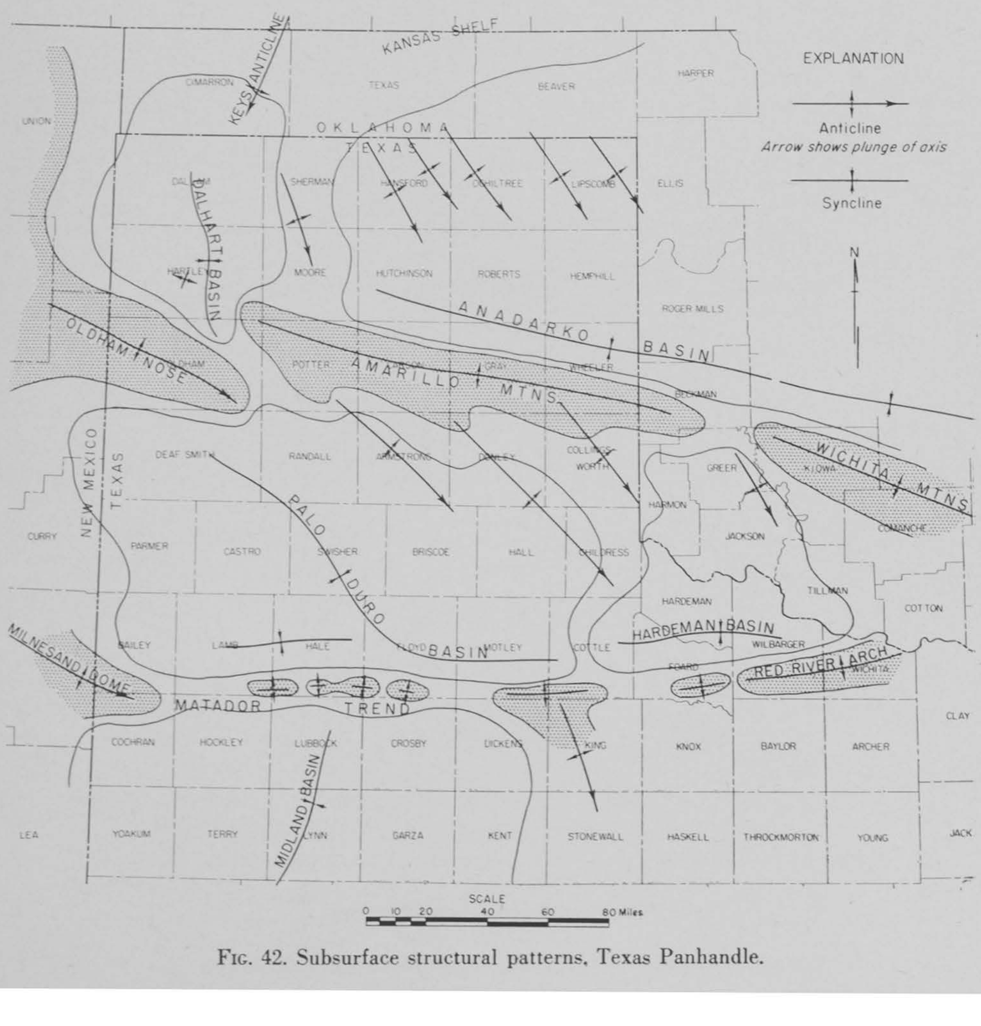


FIG. 42. Subsurface structural patterns, Texas Panhandle.

rillo Mountains, and too few wells have been drilled to establish structural patterns in that area. Northward around the north flank of the Anadarko basin, control is sufficient to establish the same general trend of folds as established south of the Amarillo Mountain trend. This pattern of folding striking obliquely to the main axis of the Amarillo uplift has been explained as secondary folding resulting from shear movement along the Amarillo Mountain front. Such a shear movement would not only explain the secondary system of folds which are readily apparent but would also explain another minor folding trend which has been discovered in the Anadarko basin paralleling the Amarillo Mountains. If shear movement occurred, this set of lesser folds would be third-order folds caused by the shear movement. Also, there is evidence of cross faulting in the Amarillo Mountains, and there is a system of parallel faults which is present along most of the south flank but is of less magnitude than the main fault on the north flank.

The Matador arch is not a continuous uplift such as the Amarillo Mountains but is a series of isolated structural peaks which are bounded on the north and south by faults which parallel the trend, and

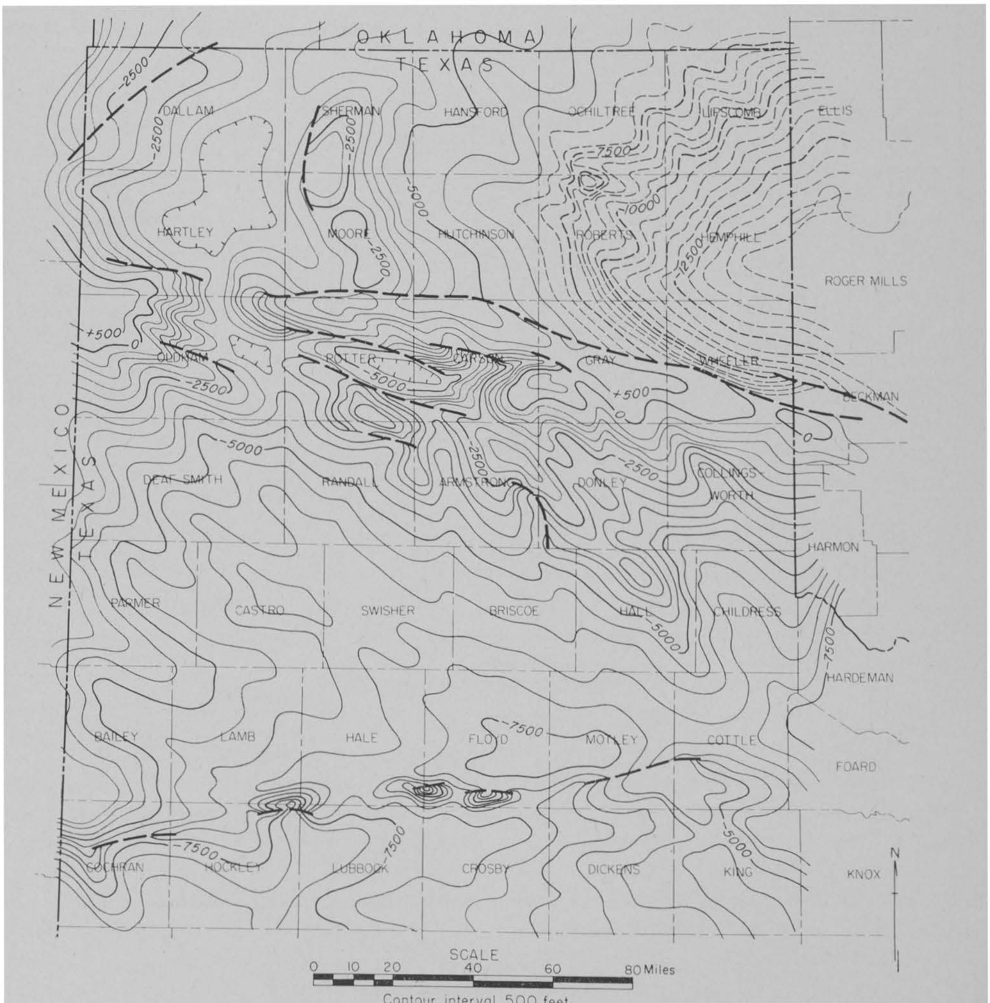


FIG. 43. Structure contour map of Precambrian basement surface.

probably by some cross faulting of a lesser magnitude. This system also has certain characteristics suggesting shear movement. The individual peaks are of varying heights, have been differentially eroded, and have some contiguous folds extending outward in a pattern similar to that associated with the Amarillo uplift. The exact nature of the main folding in the Matador system can only be speculated on at this time due to insufficient control.

A structural map contoured on the Precambrian basement surface shows the very prominent Anadarko basin extending into Texas from the east, the shallow Dalhart basin in the northwest part of the Texas Panhandle, and the broad, comparatively shallow Palo Duro basin extending through the southern part of the Panhandle (fig. 43). This basement map indicates very few faults; however, faulting in the area is probably much more complex and faults are more numerous than presently recognized. All faulting appears to be normal except that in some places the main fault bounding the Amarillo Mountain uplift on the north appears locally to be slightly reverse. The same fault system eastward along the front of the Wichita Mountains appears to be overthrust. It is questionable

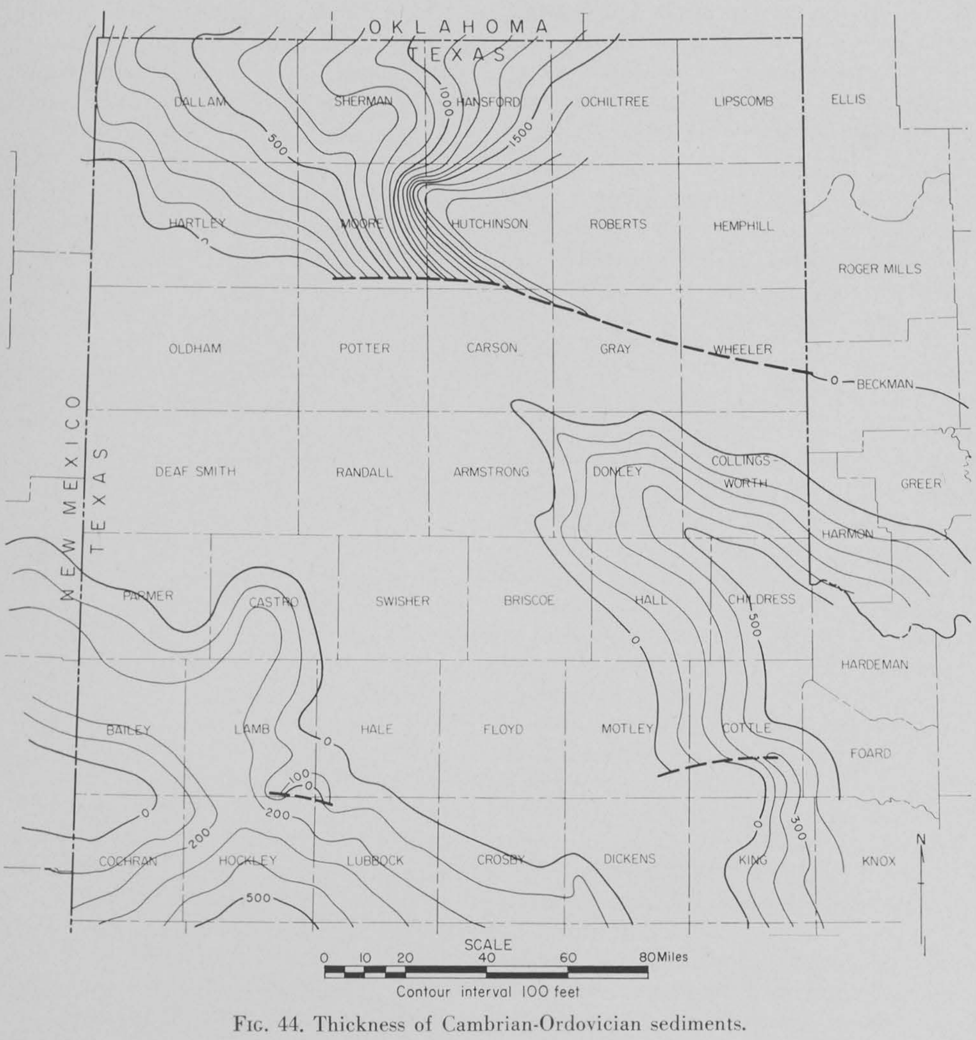


FIG. 44. Thickness of Cambrian-Ordovician sediments.

whether the overhang occurring along this fault is the result of thrusting or local high-angle reverse faulting associated with complex right lateral shearing. Since all other structural evidence suggests shearing rather than direct compression, overhangs along the fault probably resulted from second and third-order stresses along a major shear zone.

The oldest recognized structure in the area was a broad arch or warp that extended in a northwest-southeast direction across the entire Panhandle area into central Texas. This feature, named the Texas Peninsula by Adams (1954, p. 73), is dated as Lower Ordovician and is evidenced by the erosion of the Cambro-Ordovician sediments along its flanks. The position of this feature is demonstrated by isopaching the pre-Mississippian sediments which generally define its limits. By early Mississippian time, this arch ceased to be a positive element and younger Paleozoic beds were deposited across it (fig. 44).

The Texas Panhandle basins were initiated in conjunction with the Amarillo Mountains and associated folding. For many years the age of the Amarillo Mountain uplift and other prominent structures in the Panhandle area was assumed to be mid-Dornick Hills, which is an archaic Oklahoma term equivalent to early Strawn or Des Moines. More recent evidence indicates that structural movement along the same general axes was initiated probably as early as late Mississippian time. Present structure in the area appears to have had an intermittent history throughout Paleozoic time with periods of uplift followed by periods of stability and then rejuvenated uplift along the same axes.

# STRATIGRAPHY

An epi-continental sea covered the Panhandle area through most of the Paleozoic with shore lines far removed to the north and west. Individual structural elements were exposed and eroded to sea level at various times. The Texas Peninsula was a low-lying positive element between Ordovician and Mississippian time, and the Amarillo Mountains, Matador Peaks, and associated structures were exposed and had their maximum erosion during early Strawn time. There were other lesser periods of widespread erosion over the entire region.

The stratigraphic column (fig. 45) has been simplified. The Panhandle is plagued with a variety of stratigraphic names derived from the Mid-Continent region, the Rocky Mountains, west Texas, north and central Texas, and many terms unique to this area. Usage of these terms varies. In the stratigraphic chart, the Panhandle is divided into the western Anadarko basin and the Palo Duro and Dalhart basins.

Deposits in the western Anadarko basin range in age from Upper Cambrian through the Permian with only two gaps in deposition. The Devonian is missing except for the Hunton limestone of Silurian and Lower Devonian age, and the early Pennsylvanian Springer series has not been identified in the Panhandle.

In the Dalhart and Palo Duro basins a basal sandstone, possibly the Hickory formation in part, was erratically deposited. Remnants of the Ellenburger group are shown on the isopachous map of the Cambrian-Ordovician (fig. 44). No post-Ellenburger, pre-Mississippian deposits have been recognized in these basins. The Pennsylvanian Springer series is missing, and the Morrow series is present only in the northern part of the Dalhart basin. The apparent absence of Morrow strata in the rest of the Palo Duro and Dalhart basins may be due to erroneous dating of rocks included in the Bend series. It is possible that the lower part of the Bend series in the Palo Duro and Dalhart basins is equivalent to the Morrow sediments of the western Anadarko basin.

In the Panhandle area continuous deposition occurred throughout most of the Permian. Erosion of the Upper Permian in the eastern Palo Duro basin locally removed part of the Whitehorse group, but over the remainder of the Panhandle the entire Permian except for the Ochoa series is present. Post-Paleozoic rocks include the Triassic Dockum group, the upper Tertiary Ogallala formation, and Quaternary alluvium.

The simplest way to discuss the stratigraphy of the Panhandle area is to divide the depositional sequence into three major subdivisions, each of which consists of a distinctive suite of closely related rock types (Pls. I-III): (1) the pre-Pennsylvanian sequence, which includes all sediments deposited prior to the formation of the present structural pattern; (2) the Pennsylvanian-Permian sequence, which was deposited during the growth and burial of the present structure; and (3) the post-Permian sequence.

## Pre-Pennsylvanian Sequence

The pre-Pennsylvanian consists of shelf deposits except for some late Mississippian beds in the Anadarko basin (fig. 46). The earliest unit deposited on the eroded Precambrian surface was a discontinuous sandstone which is extremely variable in thickness, composition, and texture where encountered in deep wells. It is porous and coarse to fine grained with varying amounts of glauconite. Unquestionably this deposit was derived from the eroded igneous surface and locally reworked. Its thickness varies from a few to 350 feet. This basal sandstone is overlain by beds ranging in age from Cambro-Ordovician to Mississippian. The exact age of this unit has not been determined but is inferred to be Cambrian.

Overlying this basal sandstone, and in many places deposited on the Precambrian igneous surface, is a carbonate sequence of

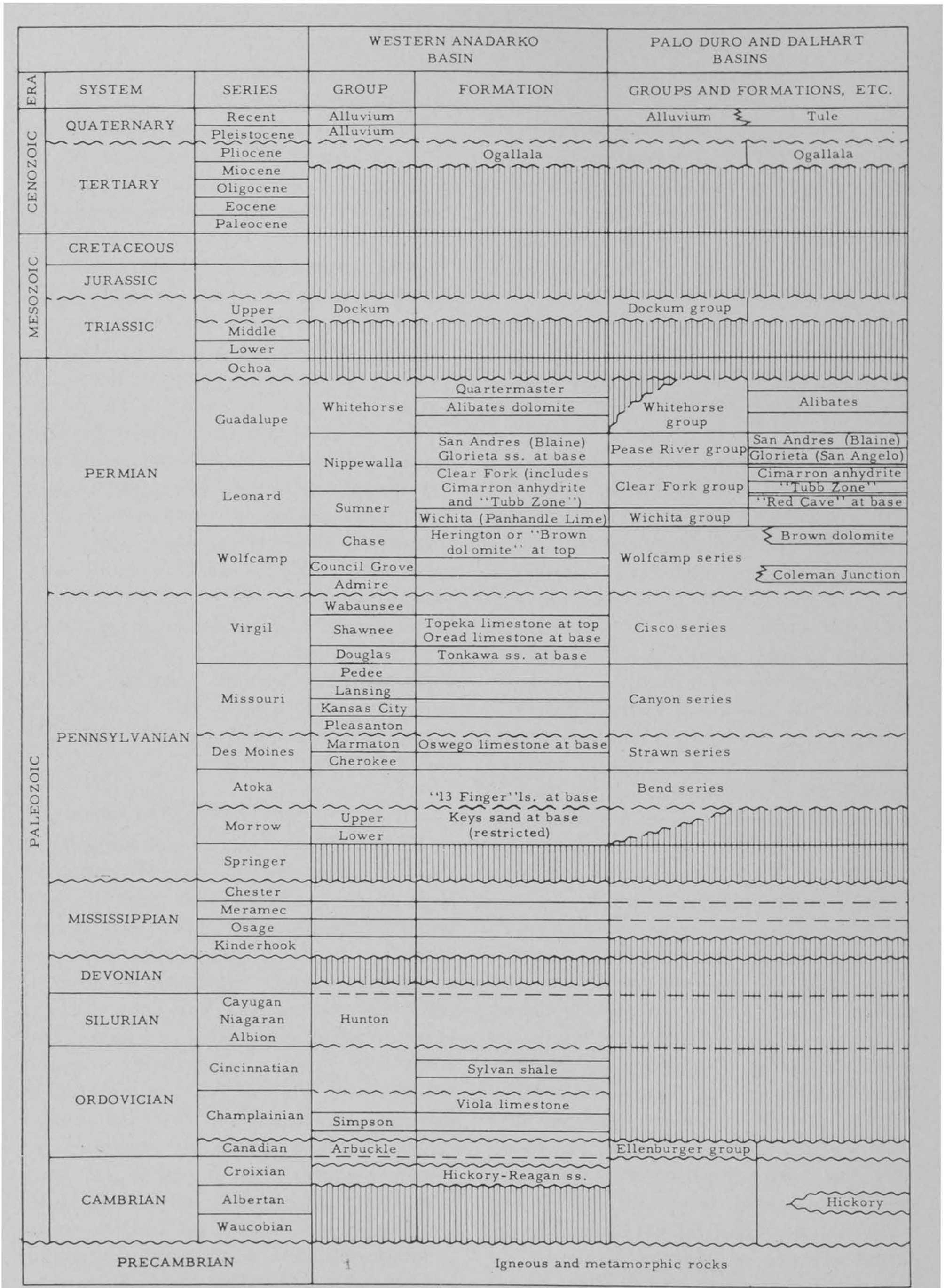


FIG. 45. Stratigraphic names used in this paper.

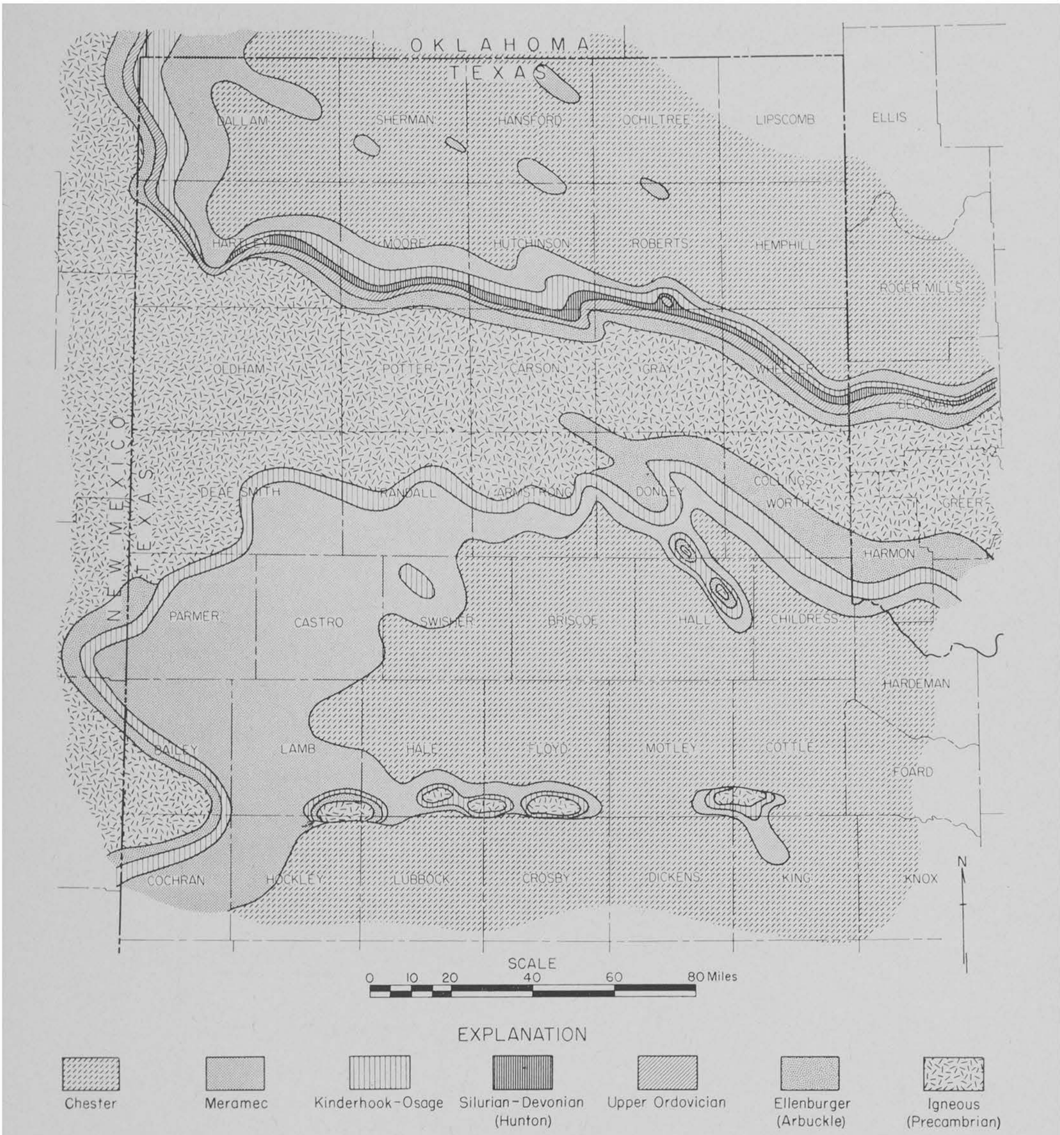


FIG. 46. Pennsylvanian subcrop map.

pre-Pennsylvanian age which is mostly limestone and dolomite with thin interbedded shales and sandstones. This pre-Pennsylvanian carbonate sequence is similar to that throughout west Texas in that it contains a number of thin chert zones. The Arbuckle or Ellenburger dolomite which occurs at the base of this interval has a maximum thickness in the area of more than 1,500 feet. The Upper Ordovician, Silurian, and Devonian is a thin sequence including the Simpson shale and limestone, the Viola limestone, the Sylvan shale, and the Hunton limestone, which has a maximum thickness of more than 800 feet. Overlying the Hunton group is a Mississippian sequence with a maximum thickness of 2,700 feet consisting of the Kinderhook sandstone, limestone, and dolomite; the Osage limestone and dolomite; the Meramec limestone and dolomite; and the Chester limestone, dolomite, sandstone, and shale.

Except for a few dark shales and thin limestones, these pre-Pennsylvanian sediments are light-colored, widespread, uniform beds interpreted as shallow-shelf deposits. This group of beds has a maximum thickness of approximately 5,000 feet in areas where the total thickness of the sedimentary sequence is more than 20,000 feet.

### Pennsylvanian-Permian Sequence

More than 70 percent of the deposits in the area are of Pennsylvanian-Permian age. The Pennsylvanian deposits were laid down while the major structures were forming and the local basins were deepening, and the overlying Permian deposits were largely formed after the principal structural growth (fig. 47). The Pennsylvanian deposits have frequent lateral and vertical facies changes which make correlations difficult over distances in some places of less than a mile. As an example, one time-stratigraphic unit, such as the Strawn series, will be found to be 100 percent granite wash near the flanks of the major uplifts. This facies a short distance away from the uplift changes to shale,

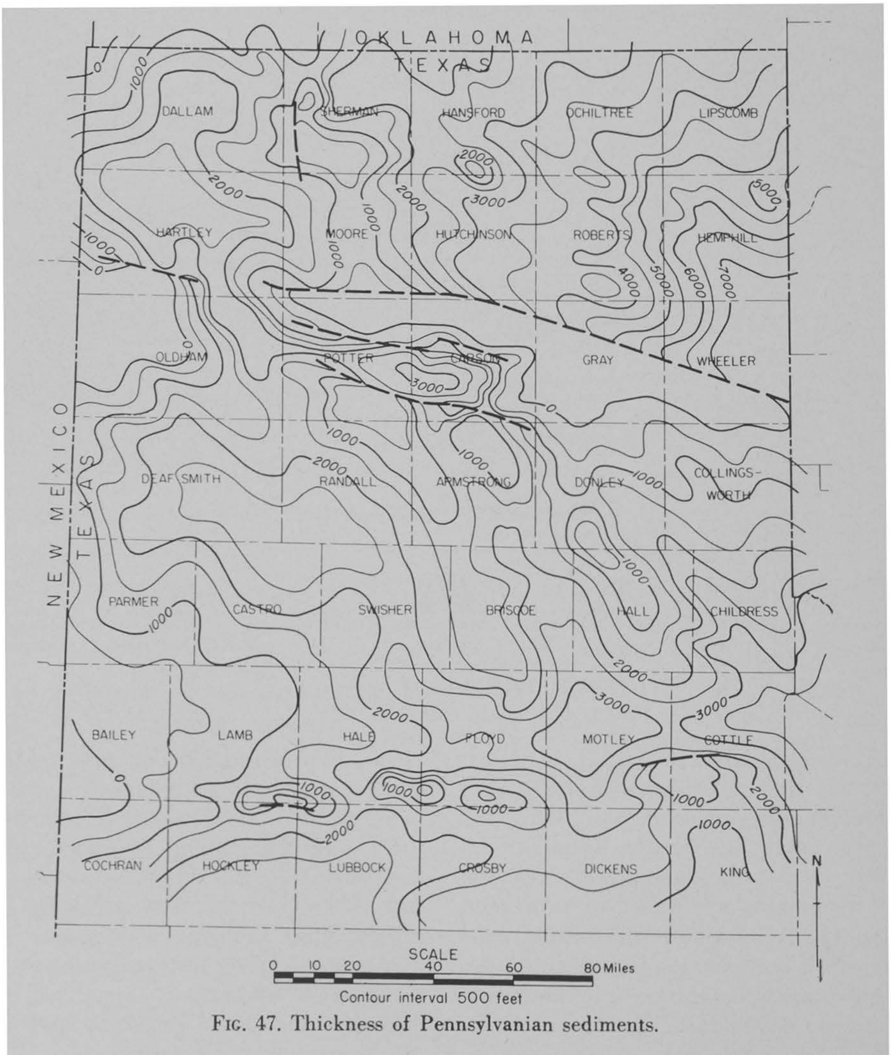


FIG. 47. Thickness of Pennsylvanian sediments.

arkose, and sandstone. Limestones, shales, and sandstones, more typically marine, occur basinward from the uplifts and associated clastic debris. This same pattern generally holds true for all younger Pennsylvanian beds. Facies changes in the early Permian are similar but less abrupt.

By Permian time the deep structural basins and channels were filled. The basins were slowly subsiding, deposition was more gradual, and the depth of water had decreased considerably. By late Permian time the local basins were entirely filled, and the uppermost beds consist of evaporitic dolomite, anhydrite, and salt with interbedded red and green shales (fig. 48).

The Pennsylvanian-Permian subdivision has been divided into six units for purposes of discussion: (1) Morrow and Atoka/Bend series, (2) Strawn/Des Moines series, (3) Canyon/Missouri series, (4) Cisco/Virgil series, (5) Wolfcamp series, and (6) post-Wolfcamp Permian.

*Morrow and Atoka/Bend series.*—The earliest identified Pennsylvanian deposits are of Morrow and Atoka age in the northern Panhandle and Atoka/Bend age in the southern Panhandle. During the time these beds were being deposited, positive areas

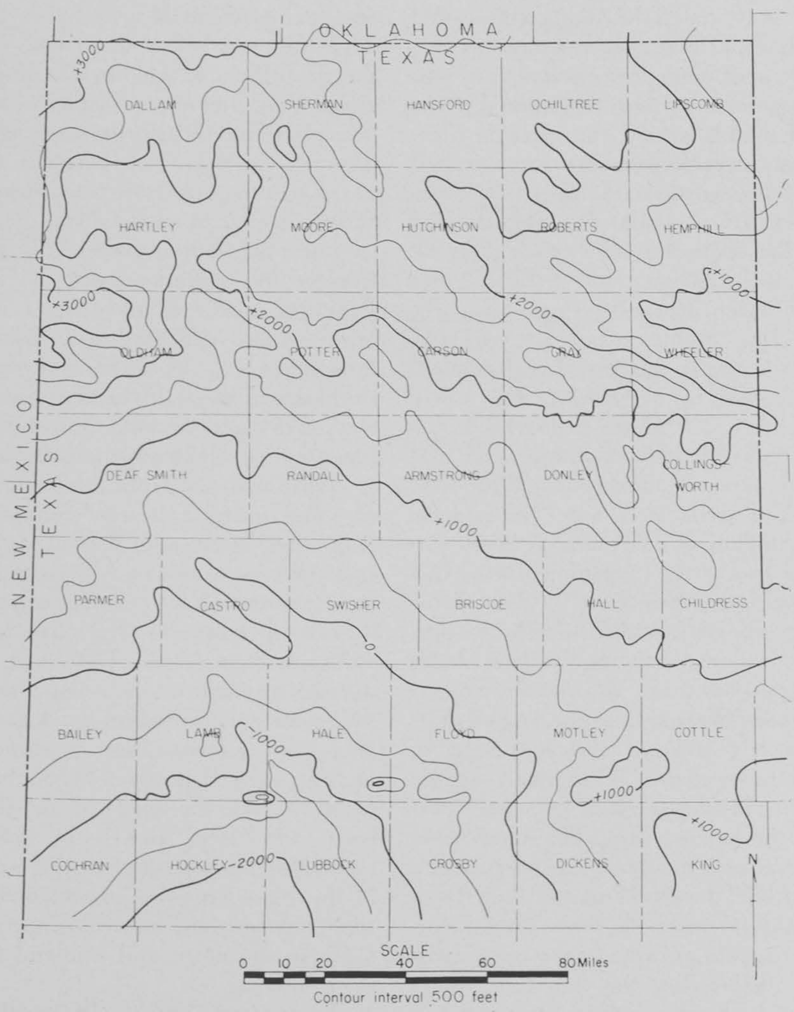


FIG. 48. Structure contour map, base of San Andres (Blaine) formation.

were rising and the basins were slowly subsiding. In the Anadarko basin the first granite washes were deposited in deep water near the flank of the Amarillo Mountain uplift and graded northward to shales and thin sandstones of the stable Kansas shelf area on the north flank of the basin. These shales and sandstones either were transported around the flank or were derived from sources farther north and west. The Palo Duro basin was shallower with more gently sloping flanks and was farther removed from the clastic source. As a result, the granite wash deposits in this basin are much more widespread and better sorted. Also during this period the Dalhart basin near the Oldham nose and the Amarillo Mountains was more or less a clastic sink, and deposits had little opportunity to be sorted and reworked. Near the end of this period considerable limestone was deposited on the Kansas shelf area on the northern flank of the Anadarko basin, and similar deposits extended across the shallow center portion of the Palo Duro basin.

*Strawn/Des Moines series.*—In early Strawn/Des Moines time the greatest amount of clastic material was deposited. These deposits in the Dalhart, Palo Duro, and southern part of the Anadarko basins consist predominantly of granite wash and arkose with minor amounts of interbedded shales and dark, deep-water limestones. The north flank of the Anadarko basin was the only area of the Panhandle during this time that remained free of the coarse clastics, and, instead, shelf limestones, thin clean bar sands, and gray marine shales were deposited.

By upper Strawn time the local uplifts were largely reduced and the flood of coarse clastics abated. Widespread organic limestone was deposited in the center portion of the basins. These limestones were biostromal with the thickest occurring in a zone around the flanks of the Palo Duro basin where conditions were most favorable to organic growth. In the center portion of this basin, the limestones were darker colored, thinner, and less organic. In the Dalhart basin the limestones developed as a discontinuous marginal deposit far removed from clastic sources. Near the Oldham nose and the Sierra Grande uplift on the west, and near the Amarillo Mountains on the south and east, a considerable amount of clastic material was still being deposited. In the Anadarko basin the deposition of the limestones was restricted to the north flank far from the immediate Amarillo Mountain clastic source.

*Canyon/Missouri series.*—The carbonate shelves which were developing at the close of Strawn time continued to develop during Canyon time. The center portions of the basins appear to have remained deep, and very little mid-basinal limestone was deposited. Marine life was prolific and a massive limestone sequence was deposited along the north, east, and west flanks of the Palo Duro basin. In the Dalhart basin the shelf area broadened after Strawn time but clastics still dominated near the old source areas. This was also true for the Anadarko basin.

A very small amount of fine clastic material—shale, siltstone, and fine-grained sandstone—was deposited in the center of the basins during Canyon time (*see* Adams et al., 1951). In the Palo Duro basin, shelf limestone development blocked the influx of clastic material from the basin periphery, and most of it was trapped in the back shelf area and deposited in local structural lows. In the Anadarko basin, a large amount of fine clastic material continued to come into the basin from the north, west, and east. For short periods during Canyon time, these shales and sandstones were deposited on the Kansas shelf area and restricted the development of organic shelf limestone. Throughout this period of time the center of the Anadarko basin was deep, and there were more clastics being deposited in this area than in the Palo Duro basin, which was in a sheltered position between the Anadarko basin to the north and Midland basin to the south.

*Cisco/Virgil series.*—In early Cisco/

Virgil time the deposition remained virtually the same as during Canyon time. By late Cisco time the finer clastic material which was slowly filling the surrounding basins reached the Palo Duro basin and prevented further development of organic shelf limestone. Gray marine shales and thin fine- to medium-grained gray sandstones were widely deposited throughout all three basins.

The Anadarko and Dalhart basins were filled earlier than the Palo Duro basin. By middle Cisco time, shallow-water sandstones, thin limestones, and thick gray marine shales extended over most of the Anadarko basin and all of the Dalhart basin. The Palo Duro basin was filled near the end of Pennsylvanian time. Following the filling of the basins, a thin shallow-water biostromal limestone was developed discontinuously throughout the areas.

The basins were filled either during a stable period of basin development during which the clastic flood was able to catch up with basinal growth, or by an abnormal amount of clastics poured into the area near the end of the Pennsylvanian.

*Wolfcamp series.*—The basins again deepened during late Cisco/Virgil or early Wolfcamp time. During Wolfcamp time marine facies developed which were similar to those of Upper Pennsylvanian age. Throughout most of Wolfcamp time carbonate shelves were developed around the flanks of the basins and extended over most of the old structural uplift areas. Marine shales and sandstones were deposited in the center portions of the basins. By the end of Wolfcamp time the entire Panhandle was a site of carbonate deposition, largely dolomite rather than limestone. This is indicative of restricted marine conditions, and the carbonate deposits probably were penecontemporaneously dolomitized.

The stratigraphic development of the Pennsylvanian and early Permian depositional sequence is best shown in the central Palo Duro basin where structure is least prominent in the whole province and clastic sources were far removed in all directions. It is here that the rock units were deposited most uniformly in an undisturbed environment throughout most of the basinal cycle (Pl. IV).

*Post-Wolfcamp Permian.*—During the rest of the Permian, restricted marine conditions existed throughout the Panhandle. The dolomites of Wolfcamp time were followed by the deposition of evaporites—anhydritic dolomites, anhydrites, and salt of the Leonard and Guadalupe series. Clastics deposited in the Panhandle area during this time are predominantly terrigenous and consist of red and green shales and red sandstone. During Upper Permian, some thin beds of dolomite were deposited in less restricted marine conditions.

## Post-Permian Sequence

By the end of Permian time basin development ceased. Slight structural movement may have continued after this and, in fact, may be continuing today, but the continental mass was uplifted regionally and no marine deposits younger than Permian are known in the Panhandle area. The overlying deposits of Triassic, Tertiary, and Quaternary age are predominantly shale and sandstone with some thin-bedded, fresh-water limestones and gypsum deposits.

## REFERENCES

ADAMS, J. E. (1954) Mid-Paleozoic paleogeography of central Texas, *in* Guide Book, Cambrian Field Trip—Llano area, San Angelo Geol. Soc., March 19–20, 1954, pp. 70–73.

ADAMS, J. E., FRENZEL, H. N., RHODES, M. L., and JOHNSON, D. P. (1951) Starved Pennsylvanian Midland basin: Bull. Amer. Assoc. Petrol. Geol., vol. 35, p. 26.

GIRARD, R. M. (1959) Bibliography and index of Texas geology, 1933–1950: Univ. Texas Pub. 5910, 238 pp.

GOULD, C. N. (1907) Geology and water resources of the western portion of the Panhandle of Texas: U. S. Geol. Survey Water Supply Paper 191, 70 pp.

———— (1923) Crystalline rocks of the Plains: Bull. Geol. Soc. Amer., vol. 34, pp. 541–560.

———— (1924) A new classification of the Permian red beds of southwestern Oklahoma: Bull. Amer. Assoc. Petrol. Geol., vol. 8, pp. 322–341.

————, and LEWIS, F. E. (1926) The Permian of western Oklahoma and the Panhandle of Texas: Oklahoma Geol. Survey, Circ. 13, 29 pp.

ROGATZ, HENRY (1935) Geology of Texas Panhandle oil and gas field: Bull. Amer. Assoc. Petrol. Geol., vol. 19, pp. 1089–1109.

———— (1939) Geology of Texas Panhandle oil and gas field: Bull. Amer. Assoc. Petrol. Geol., vol. 23, pp. 983–1053.

ROTH, ROBERT (1955) Paleogeology of Panhandle of Texas: Bull. Amer. Assoc. Petrol. Geol., vol. 39, pp. 422–443.

SELLARDS, E. H. (1933) Bibliography and subject index of Texas geology, *in* The geology of Texas, Vol. I, Stratigraphy: Univ. Texas Bull. 3232 (Aug. 22, 1932), pp. 819–996.

TOTTEN, R. B. (1956) General geology and historical development, Texas and Oklahoma Panhandles: Bull. Amer. Assoc. Petrol. Geol., vol. 40, pp. 1945–1967.

# Conjectured Middle Paleozoic History of Central and West Texas[7]

JAMES LEE WILSON[8] and O. P. MAJEWSKE[9]

## ABSTRACT

Regional Siluro-Devonian faunal correlations between outcrops in the Llano uplift of Texas, the Arbuckle Mountains of Oklahoma, and the Trans-Pecos Texas—southern New Mexico mountains are made with strata of the west Texas subsurface whose faunas have recently been studied. These correlations are based upon outcrop areas generally representing tectonic shelves with thin, incomplete, but fossiliferous limestones as well as the thicker but less fossiliferous strata of marginal cratonic basins.

The major segments of middle Paleozoic strata in the southwestern states are:

(1) A widespread thin pure carbonate sequence of Lower Silurian (Alexandrian) and lower Niagaran age.

(2) A lithologically diverse and thicker unit of Middle and Upper Silurian (Niagaran) age represented by marls and thin limestones and in parts of west Texas and all of New Mexico by massive dolomitized platform-type carbonates.

(3) An unconformity above the Silurian at the position occupied by the evaporites of the Michigan and New York basins. This is present also in the west Texas and Anadarko basins.

(4) A widespread unit of fossiliferous, generally limestone, strata of Lower Devonian through Onondagan age, thin in the Ozark, Arbuckle, and Llano uplift areas but as much as 1,100 feet thick in the west Texas basin.

(5) A second major unconformity of late Middle Devonian age as widespread as the Siluro-Devonian one.

(6) A widespread sheet of relatively thin argillaceous and siliceous deposits typical of the Devono-Mississippian black shales of the Mid-Continent. These Woodford deposits grade westward into argillaceous limestones in New Mexico and west Texas.

From isopach and subcrop maps of these major stratigraphic segments the presence of the Paleozoic Texas craton may be inferred. The Kerr and Fort Worth basins on its margins are apparently post-middle Paleozoic features. Silurian beds were more restricted than Devonian and are overlapped by the latter over the Texas craton and in the eastern Arbuckle Mountains. It may be conjectured that the faunal differences between west Texas and Hunton Silurian beds are in part caused by deposition in separate basins. The older Devonian of the Southwest is correlative eastward with the Appalachian section and the seas extended into the west Texas basin, where a thick section is preserved. However, Lower and lower Middle Devonian is not present in New Mexico outcrops, and the seas probably did not extend this far west. The Woodford covered the the entire area after a Middle Devonian erosion period.

## INTRODUCTION

Middle Paleozoic strata on and marginal to the North American craton are difficult to study because they occur in thin, scattered outcrops over the positive areas and are relatively inaccessible where more fully developed in the basins. Despite such difficulty, this paper attempts some tenta-

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tive general conclusions, based on recent biostratigraphic studies in both subsurface and outcrop localities, about correlation and historical interpretation of these strata in the southwestern states. The main areas of interest are the west Texas subsurface basin, the outcrops in southern New Mexico and Trans-Pecos Texas, and the Llano uplift; also of importance are outcrops in the Ouachita and Arbuckle Mountains of southern Oklahoma and the subsurface of the Anadarko basin. These areas encompass two types of depositional provinces: (1) ancient positive elements with fossiliferous shelly limestones but thin and interrupted sections and (2) basins marginal to the craton with thicker, less fossiliferous sections. For a complete understanding of regional history, it is important that the geologic column in both types of provinces be considered. Recognition of biostratigraphic units is generally possible in the fossiliferous beds of such areas as the Llano uplift and Ozark dome, but evaluation of regional unconformities and recognition of widely persistent lithic units can be accomplished only when the stratigraphy and a degree of faunal zonation have also been worked out in the basinal areas.

The following main sources of information were utilized in preparing this paper:

(1) Study of cores from 33 wells in west Texas, with additional investigations by the writers of outcrops in New Mexico and Trans-Pecos Texas.

(2) Recent papers by Thomas Amsden, of the Oklahoma Geological Survey, describing both fauna and stratigraphy of the Hunton group of the Arbuckle Mountains.

(3) Studies of stratigraphy and shelly faunas of the Llano uplift Devonian remnants made recently by P. E. Cloud, Jr., and V. E. Barnes, U. S. Geological Survey and Bureau of Economic Geology, respectively.

(4) Conodont studies of the Devonian black shale by W. H. Hass, U. S. Geological Survey; S. P. Ellison, The University of Texas; K. J. Müller, Berlin Technical Institute; and R. W. Graves, The California Company.

In this paper the paleontological work on the Silurian and Devonian of west Texas is that of O. P. Majewske, and the regional synthesis is by J. L. Wilson.

## OUTLINE OF BIOSTRATIGRAPHY IN AREAS OF STUDY

The correlation chart (fig. 49) shows the age relations of the Silurian and Devonian in eight key areas. The middle Paleozoic beds include three major sequences of strata: The highest sequence consists of dark argillaceous beds of Upper Devonian[10] age; the middle is a low Middle and Lower Devonian sequence of carbonate and chert, separated by unconformity from the overlying Upper Devonian and also from the underlying Silurian; and the lowest sequence, the Silurian, consists of two unequal parts—the thick upper portion is referred to the widely recognized Niagaran of the Mid-Continent region, the thin persistent unit below is referred to the Alexandrian.

Many fossils of value for correlation and used as a basis of conclusions in this paper are mentioned in the sections which follow. However, only a limited number of the better specimens from outcrops and cores in west Texas studied by Majewske are shown on Plates V and VI. The material shown is fairly representative of the Silurian and older Devonian, as indicated in the plate explanations, but is not a complete representation of the fauna, which contains many corals and bryozoans not illustrated.

### ARBUCKLE MOUNTAINS AND ANADARKO BASIN, HUNTON GROUP AND WOODFORD FORMATION

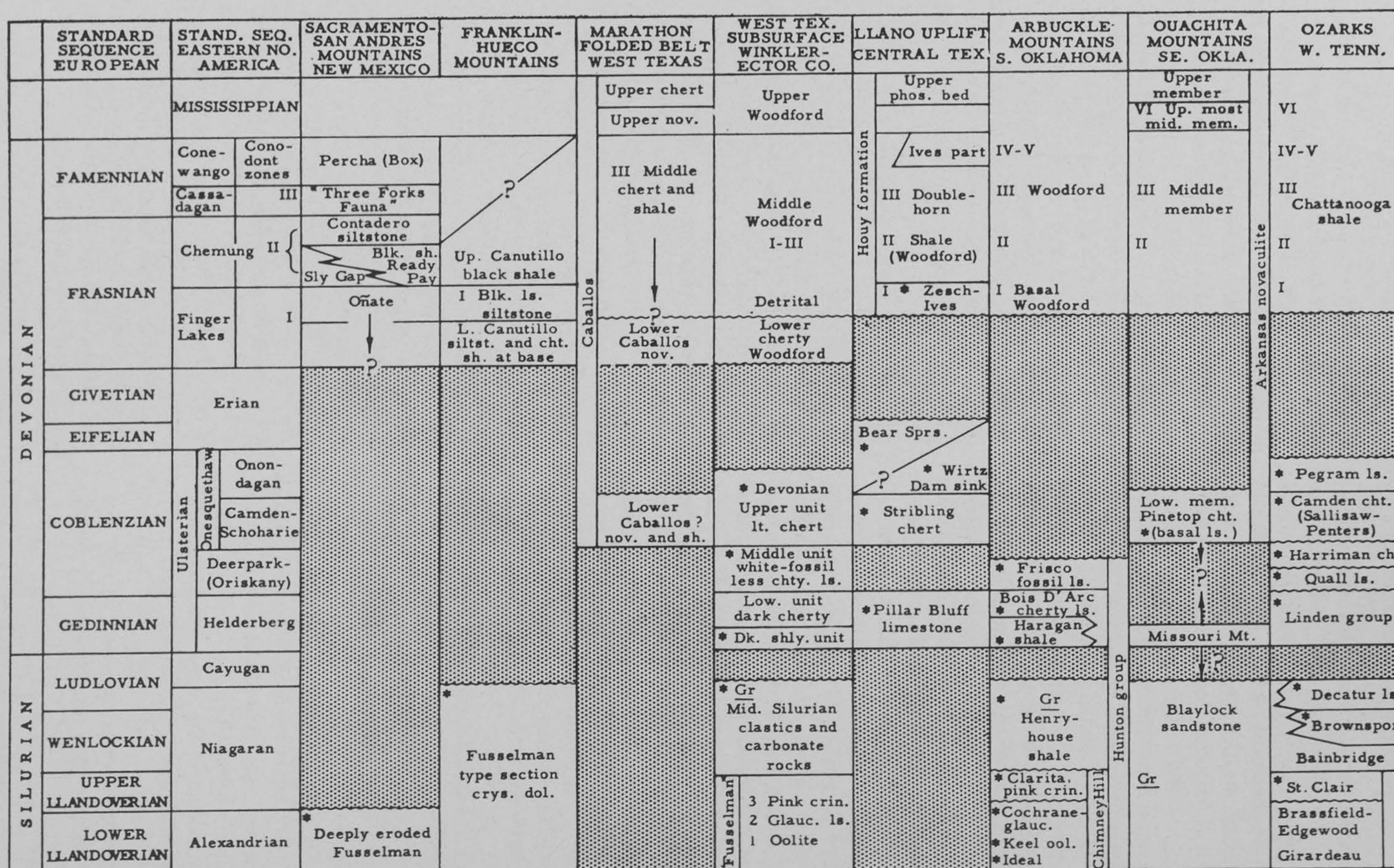
Thanks to W. H. Hass (1956a) a well-systematized Late Devonian conodont sequence is now known from the Woodford which may be correlated accurately with the Chattanooga shale of the Ozark Mountains and western Tennessee. The Woodford formation in the Arbuckle Mountains is predominantly varicolored chert interstratified with dark siliceous shale; it reaches a maximum thickness of 600 feet. The formation lies unconformably on the Hunton group, mainly resting on its upper beds which are of Early Devonian age. These in turn consist of the Frisco formation, a thin (40 feet, maximum) cherty fossiliferous limestone of Deerparkian age at the top, and the Bois d'Arc formation, a marl and calcarenite as much as 200 feet thick lying disconformably below. At least part of the Bois d'Arc in the Arbuckles is a facies of the Haragan shale (Amsden, 1957, pp. 43–44 and fig. 4). The faunas of the Bois d'Arc-Haragan units have been described and illustrated by Amsden and Boucot (1958) who confirm their established Helderbergian age. A faunal hiatus occurs within the Hunton group of the Arbuckle Mountains separating the Haragan shales from the Silurian Henryhouse formation, an almost identical marly limestone and shale unit also as much as 200 feet thick. The Henryhouse fauna is correlated with the upper Niagaran Brownsport formation of western Tennessee (Amsden, 1951, pp. 70–71) and is also similar to that of the overlying Haragan. Both the Henryhouse and Haragan represent the same biofacies, and the two faunas are probably not far removed in time. Nevertheless, the faunas may be clearly distinguished by detailed paleontology, and, although lithically indistinct in a normal section, the unconformity between the Haragan and Henryhouse is important, for in places the entire Henryhouse formation has been removed by pre-Devonian erosion (Amsden, 1957, p. 31).

In addition to brachiopods, corals, trilobites, and bryozoans, the Henryhouse shale contains graptolites which have been identified by Decker (1935, 1936) with forms from the uppermost Silurian (Ludlovian) of the British Isles.

The Chimneyhill formation lies disconformably below the Henryhouse formation and consists of four members, in descending order (Amsden, 1957):

Clarita member—a thin (30 feet thick) widespread crinoidal limestone;
Cochrane member—a glauconitic limestone

[10] The writers use of the terms Upper, Middle, and Lower Devonian is that of Cooper et al. (1942). In their Devonian correlation chart, unequivocal Lower Devonian encompasses Helderbergian through Oriskanian beds of the New York section.




*Shelly faunas. I - VI, conodont zones of W. H. Hass.- Gr, graptolite faunas.

FIG. 49. Correlation chart of Silurian and Devonian in the southwestern part of the United States.

about 60 feet thick (maximum) and separated from the Clarita by an unconformity;
Keel member—an oolitic limestone 0 to 15 feet thick, at places divided into two oolitic zones by calcilutite;
Ideal Quarry member — brown-weathering, cherty calcarenite, 3 to 5 feet thick, disconformably overlying the Ordovician Sylvan shale.

The Clarita member is faunally and lithically equivalent to the Niagaran St. Clair limestone of the Ozarks and part of the subsurface Fusselman of west Texas. The Cochrane, Keel, and Ideal Quarry members are equivalent to the Alexandrian Brassfield, Edgewood, and Girardeau formations of the Ozarks.

In the Anadarko basin the Hunton cannot be separated conveniently into its Silurian and Devonian parts and is generally mapped as a moderately thick carbonate unit with a central marly portion (Haragan plus Henryhouse). The unit ranges in thickness from less than 100 feet on the tectonically positive area on the north and east sides of the basin to perhaps 900 feet along the southern downwarped edge of the basin just north of the Wichita—Criner Hills axis.

The Lower Devonian portion of the Hunton of Oklahoma is correlative to a group of carbonate and chert formations in western Tennessee, faunas from which have been known since 1919 through work of C. O. Dunbar (1919). Lower Middle Devonian beds occurring above Lower Devonian in other outcropping areas of the southwestern states are not present in the Arbuckle Mountains, possibly indicating extensive truncation at the top of the Hunton group.

## Llano Uplift Section of Central Texas

No Silurian is known to occur in the Llano uplift (located at the southeastern edge of the Texas craton), but numerous remnants of Devonian units are present. This area, in contrast to the Arbuckle Mountains and Anadarko basin, was a more positive element throughout middle Paleozoic time and contains a thinner, more interrupted section. The Devonian remnant formations are generally fossiliferous, light-colored carbonate rocks representing various biostratigraphic portions of the system. Most of the units are preserved in sink holes in the Ellenburger terrane, either as deposits or as collapsed remnants preserved from later erosion. Stratigraphic and structural relations of these remnants are being worked out, mainly by Cloud, Barnes, and Hass, who have previously (1957) discussed these formations. Several periods of collapse and sink-hole deposition have occurred between the extensive pre-Devonian truncation of the Ellenburger surface in this region and deposition of the Late Mississippian Barnett shale.

The Upper Devonian of the Llano uplift consists of a remnant of the black Woodford shale (Doublehorn shale) and conodont-bearing residual chert breccias. Only one of the breccias, the Zesch unit, which has recently been equated with the older-named Ives breccia, contains a shelly fauna. Brachiopods from the Zesch in one of the large sinks (Bear Springs in Mason County, western Llano uplift) prove its Late Devonian age. Because of the work by Hass (*in* Cloud, Barnes, and Hass, 1957), the lower Woodford-Chattanooga zones can be recognized in the Doublehorn shale.

Study is still proceeding on the lower Middle and Lower Devonian portions of the remnant formations. Cloud and Barnes have identified a lower Erian (Marcellus or Couvinian) fauna from the Bear Springs sink (Cloud, P. E., personal communication, July 1958), but this fauna is somewhat younger than any other pre-Woodford Devonian known in the southwestern states and may actually prove upon further study to be of Onondagan age. A recognized Onondagan fauna has been found in the Wirtz Dam sink near Marble Falls on the eastern edge of the Llano uplift. This fauna is under study by Cloud, who at present believes that it correlates with the European Emsian (upper Coblenzian) (Cloud et al., 1957, p. 808, and personal

## PLATE V

Silurian Fossils From Wells in Various Texas Counties

FIGURES—

1, 2. *Conocardium* sp., x4. Depth 10,895 feet, Atlantic Refining Company No. 9-CE-C-1 University, Andrews County.
Figure 1. Right lateral view.
Figure 2. Anterior view.

3, 4. *Cyclospira/Protozeuga* sp., x4. Depth 10,820 feet, Gulf Oil Corporation No. 1 McElroy-State, Upton County.
Figure 3. Pedicle exterior.
Figure 4. Brachial exterior of a different specimen.

5, 6. *Coelospira* sp., x4. Depth 12,110–12,115 feet, Humble Oil & Refining Company No. 1 Weaver, Dawson County.
Figure 5. Pedicle exterior.
Figure 6. Brachial exterior of a different specimen.

7. *Triplesia* sp., x4. Brachial view. Depth 10,870 feet, Atlantic Refining Company No. 9-CE-C-1 University, Andrews County.

8. *Triplesia* sp., x4. Pedicle view. Depth 9,501 feet, Wilshire Oil Company of Texas No. 34–98 Jacobs Livestock Co., Upton County.

9. "*Clorinda*" sp., x4. Pedicle view. Depth 10,814 feet, Gulf Oil Corporation No. 1 McElroy-State, Upton County.

10. *Rhynchotreta* sp., x4. Pedicle view of broken specimen. Depth 8,905 feet, Magnolia Petroleum Company No. 25-E Walton, Winkler County.

11. *Brachymimulus* sp., x4. Pedicle view of incomplete specimen. Depth 9,596 feet, Wilshire Oil Company of Texas No. 34–98 Jacobs Livestock Co., Upton County.

12. *Plectatrypa* sp., x4. Pedicle view. Depth 8,894 feet, Magnolia Petroleum Company No. 25-F Walton, Winkler County.

13, 14. *Dinobolus* sp., x1. Depth 12,107 feet, Humble Oil & Refining Company No. 1 Weaver, Dawson County.
Figure 13. Pedicle exterior.
Figure 14. Cast of brachial interior.

15. *Eospirifer* sp., x2. Rubber cast of pedicle valve exterior. Depth 12,087 feet, Shell Oil Company No. 1 Clay, Dawson County.

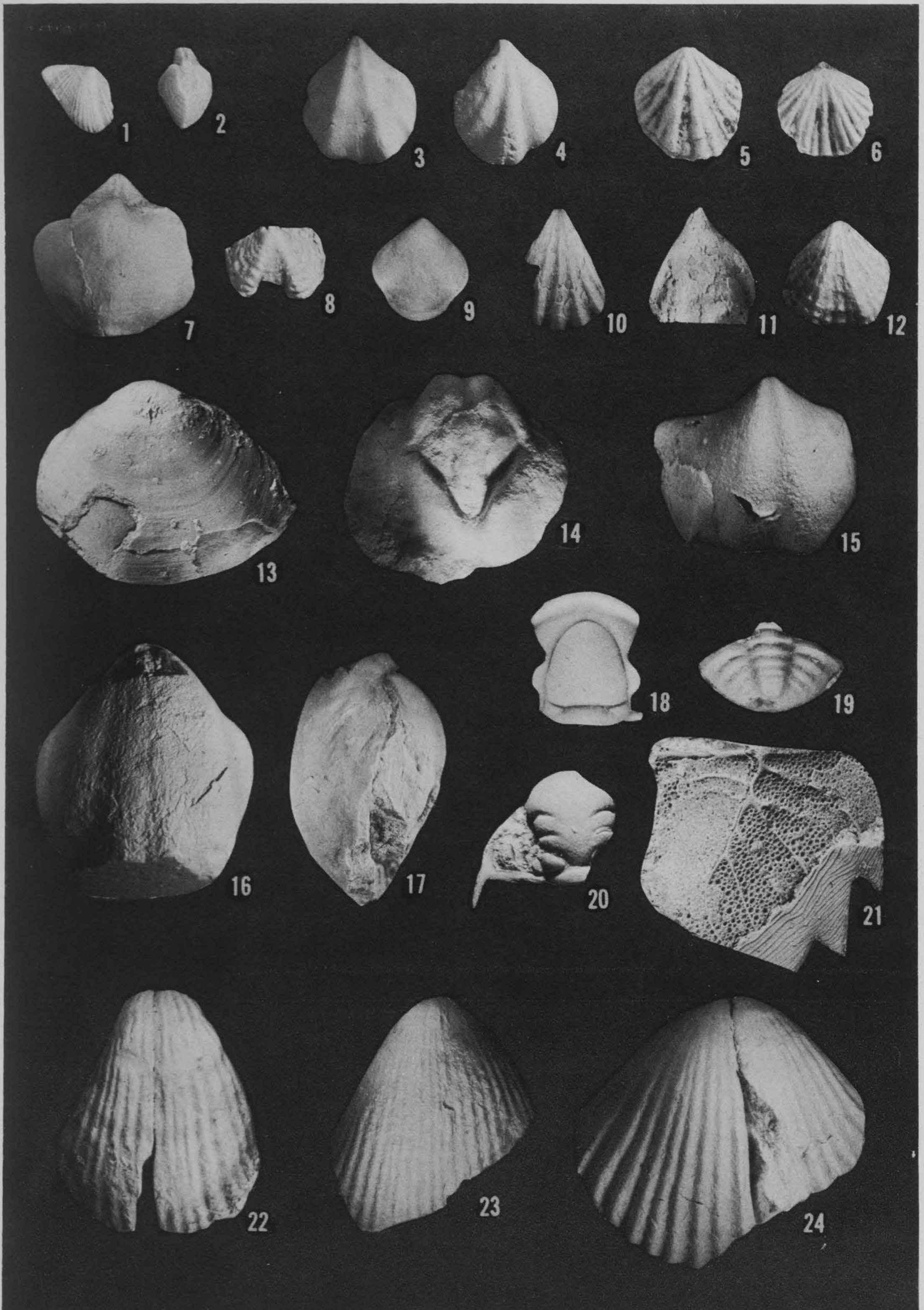
16, 17. "*Pentameroides*" sp., x2. Depth 11,891 feet, Union Oil Company of California No. 1-10 Culp, Cochran County.

18, 19. *Proetus* sp., x4. Depth 10,810 feet, Gulf Oil Corporation No. 1 McElroy-State, Upton County.

20. *Cheirurus* sp., x2. Depth 12,109 feet, Humble Oil & Refining Company No. 1 Weaver, Dawson County.

21. *Arctinurus* sp., x2. Impression of fragmentary pygidium. Depth 12,109 feet, Humble Oil & Refining Company No. 1 Weaver, Dawson County.

22–24. *Conchidium* spp., x2. Pedicle views.
Figures 22, 24. Depth 10,554 feet, Atlantic Refining Company No. 9-CE-C-1 University, Andrews County.
Figure 23. Depth 12,474 feet, Forest Oil Corporation-Monterey Oil Company No. 2-E University, Andrews County.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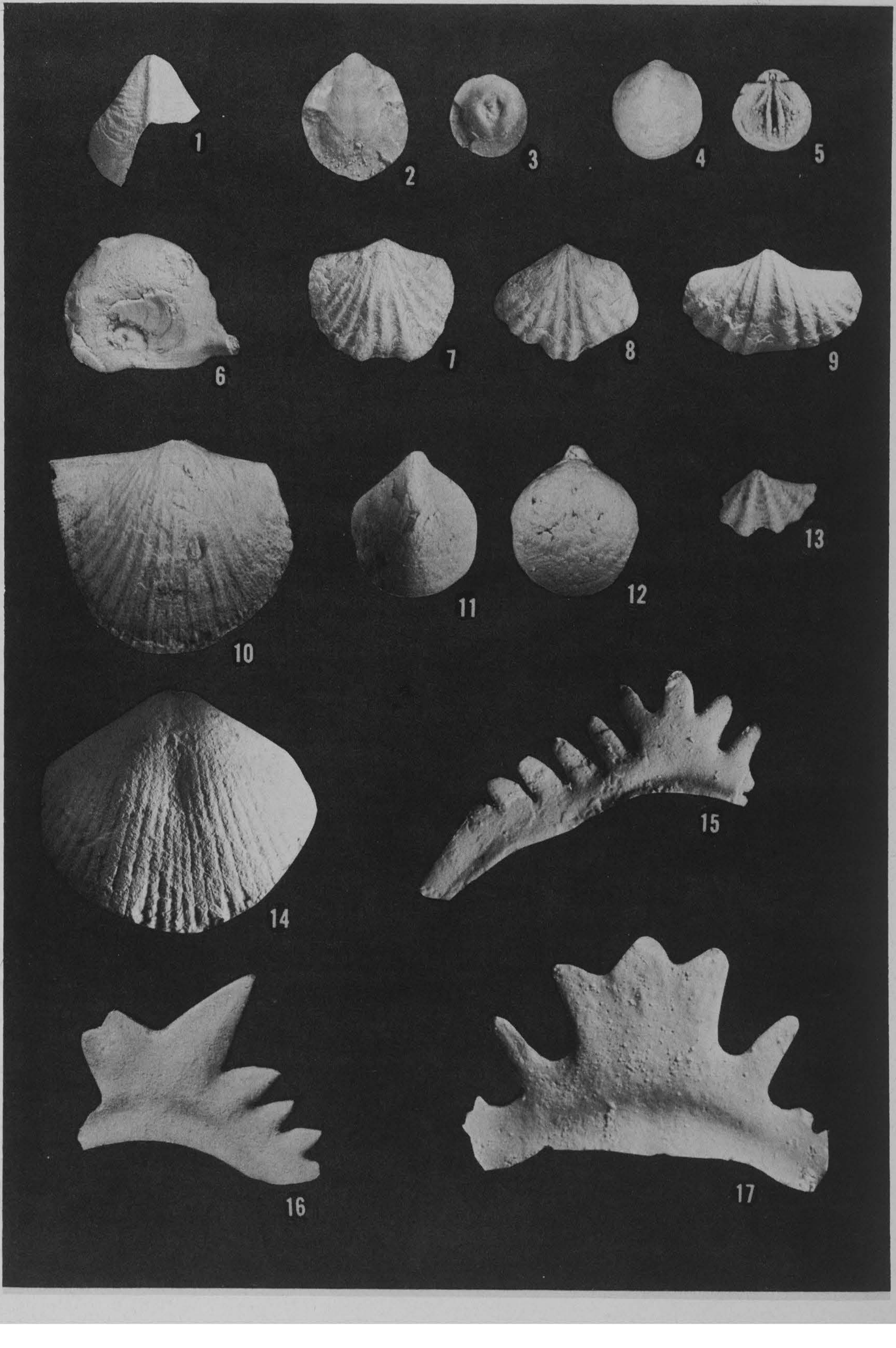
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## PLATE VI

Devonian Fossils From Wells in Various Texas Counties

FIGURES—

1. "*Crytolites*" sp., x2. Depth 10,486–10,488 feet, Buffalo Oil Company-Midstates Oil Company No. B-2 University, Andrews County.

2, 3. *Lingulapholis* sp., x4. Depth 10,796 feet, Shell Oil Company No. B-2 University, Andrews County.
Figure 2. Pedicle exterior.
Figure 3. Brachial interior.

4, 5. *Anoplia* sp., x4. Depth 11,652–11,653 feet, Forest Oil Corporation-Cities Production Company No. 46-2 Fee, Midland County.
Figure 4. Pedicle exterior.
Figure 5. Brachial interior.

6. *Platyceras* sp., x2. Side view of small exfoliated specimen. Depth 10,511–10,512 feet, Buffalo Oil Company-Midstates Oil Company No. B-2 University, Andrews County.

7, 8. *Leptocoelia* sp., x2. Depth 10,201 feet, Texas Pacific Coal & Oil Company No. B-1 Johnson, Ector County.
Figure 7. Pedicle view.
Figure 8. Brachial view of a different specimen.

9. *Delthyris* sp., x2. Pedicle valve. Depth 10,135 feet, Texas Pacific Coal & Oil Company No. B-1 Johnson, Ector County.

10. *Eodevonaria* sp., x4. Partially exfoliated pedicle valve. Depth 10,637 feet, Shell Oil Company No. B-2 University, Andrews County.

11, 12. *Centronella* sp., x4. Depth 11.656–11,657 feet, Forest Oil Corporation-Cities Production Company No. 46-2 Fee, Midland County.
Figure 11. Pedicle view.
Figure 12. Brachial view of a different specimen.

13. *Kozlowskiella* sp., x2. Fragment of pedicle valve. Depth 11,032 feet, Shell Oil Company No. D-2 University, Andrews County.

14. *Costispirifer* sp., x1. Pedicle valve. Depth 11,751–11,752 feet, Humble Oil & Refining Company No. B-1 Methodist Home, Gaines County.

15, 17. *Synphoroides* sp. Fragments of frontal processes. Shell Oil Company No. B-2 University, Andrews County.
Figure 15. x4. Depth 10,637 feet.
Figure 17. x4. Depth 10,670 feet.

16. *Synphoroides* sp., xl. Fragment of frontal processes. Depth 10,968 feet, Shell Oil Company No. E-3 University, Adrews County.

correspondence, July 1958). Brachiopods are also known from a still older remnant formation, the Stribling light-colored limestone and chert. The brachiopods date this remnant as lower Onesquethawan, equivalent to the Camden chert of Tennessee and the Schoharie formation of New York State. Most pre-Woodford units found in central Texas are younger than any Devonian known in the Hunton group. Although no Deerpark deposits (Frisco equivalent) have been recognized in the Llano uplift remnants, a Helderberg fauna similar to that of the Haragan shale has been discovered from the Pillar Bluff formation (Barnes et al., 1947, p. 129) in one or two sink holes.

With the exception of a 200-foot thick unit—which has lithic character much like the Stribling of the Llano uplift—in the Rowsey No. 2 Nowlin well in Kerr County (Barnes, 1959, cross section, Pl. 1), no Devonian has been reported from the subsurface in the vicinity of the Llano uplift.

### West Texas Subsurface Basin

In the great basinal area of west Texas (Tobosa basin of Galley, 1958), a much more complete section of the Devonian and Silurian is present. The upper part of the sequence is the Woodford formation, whose black shale and cherty beds are typical of the basinal facies of the formation over all of the Southwest. The age of the Woodford shale in west Texas as well as the Woodford and Chattanooga of Oklahoma and eastern areas has been much debated, but correlations are now fairly clear. According to Ellison (1950, p. 17), the Woodford has three members where it is fully developed in Winkler County (*see also* Jones, 1953). The lower cherty member is apparently confined to Ward and Winkler counties, and its exact lithic equivalent elsewhere is not known. Possibly it is the lower part of the Caballos novaculite and/or the cherty unit which forms the lower half of the Canutillo formation of Trans-Pecos Texas (Jones, 1953, chart on p. 16). Ellison (1950) and Ellison and Wynn (1950) recognized groups of conodont species that can now be related to the Chattanooga-Woodford conodont zones of Hass (1956a). The middle Woodford of Ellison is now known to encompass at least zones I through III and probably also the higher zones IV, V, and VI of Hass' section. The upper part of the west Texas Woodford is in all probability Mississippian by analogy with the Woodford formation in Oklahoma.

The area of greatest preserved Woodford thickness probably coincides with a depositionally negative area near the center of the west Texas pre-Pennsylvanian basin and lies northwest of the area of maximum preserved older Devonian (figs. 50 and 51). In the west Texas basin the Woodford rests with unconformity on a carbonate rock unit of older Devonian and Silurian age which generally has not been separated in previous stratigraphic studies of the area.

Separation of this extensive carbonate body into Silurian and Devonian portions is facilitated through faunal identifications by O. P. Majewske reported here from the distinctive lithic units within the "Siluro-Devonian" of Winkler, Ward, Ector, Midland, Crane, and Upton counties (Jones, 1953, p. 14). The Devonian portion of these units is as follows:

| | *Thickness in feet* |
|---|---|
| Top— | |
| Light-colored chert and limestone | 200 |
| Fossiliferous calcarenitic limestone | 450 |
| Dark chert and cherty limestone | 100–300 |
| Dark shale with conodonts and spores | 0–45 |
| Base. | |

The youngest strata within this sequence are possibly lower Middle Devonian (Onondagan) based on the occurrence in a well in Upton County of a frontal margin of a trilobite, *Odontocephalus*, about 40 feet below the Woodford and almost 900 feet above the top of the Silurian silty limestone. Numerous wells in the counties mentioned above contain brachiopods and trilobites (fig. 52) from the upper two Devonian carbonate units listed above, indicating a late Early Devonian age for the

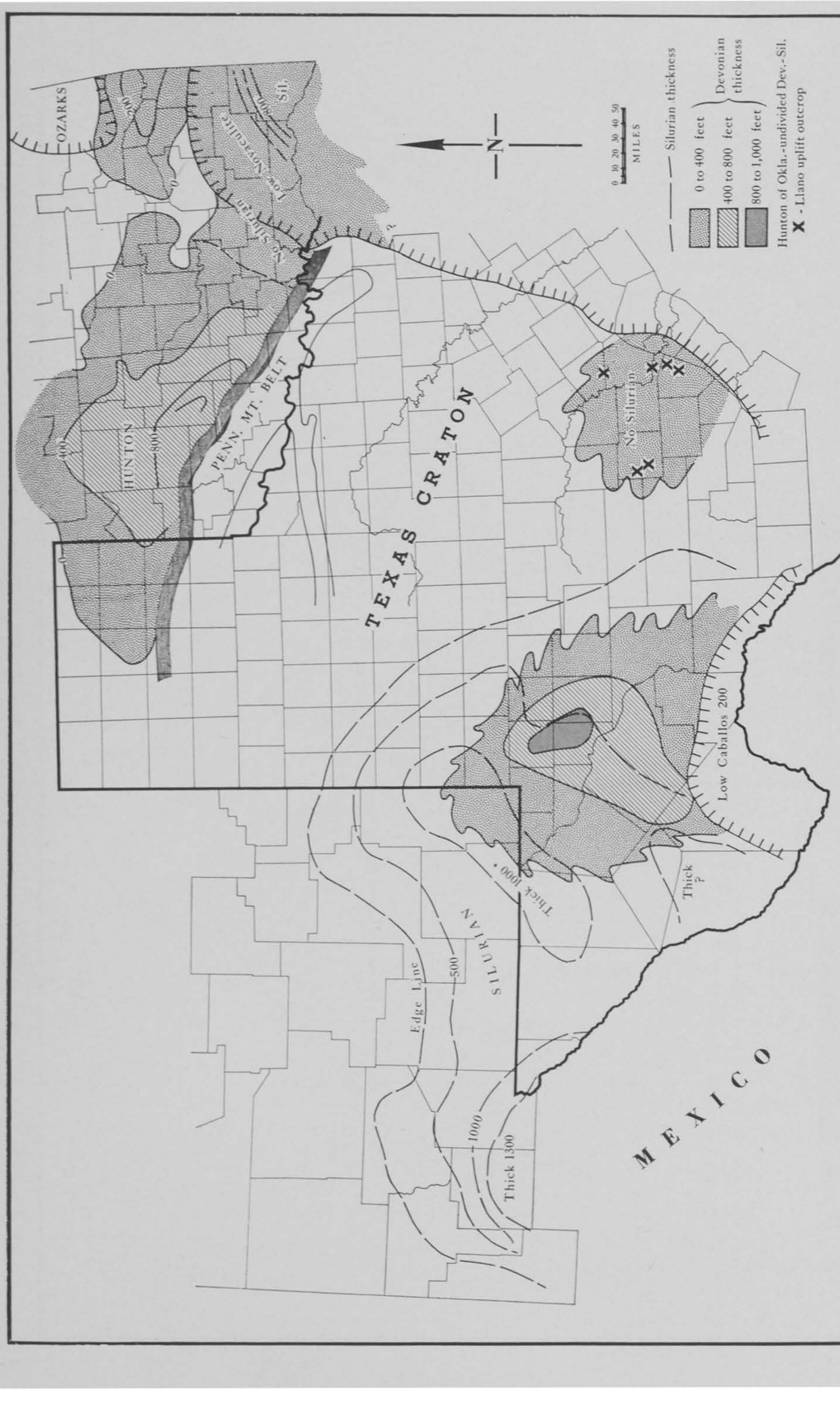


FIG. 50. Isopach map of Silurian and older Devonian strata in the Southwest.

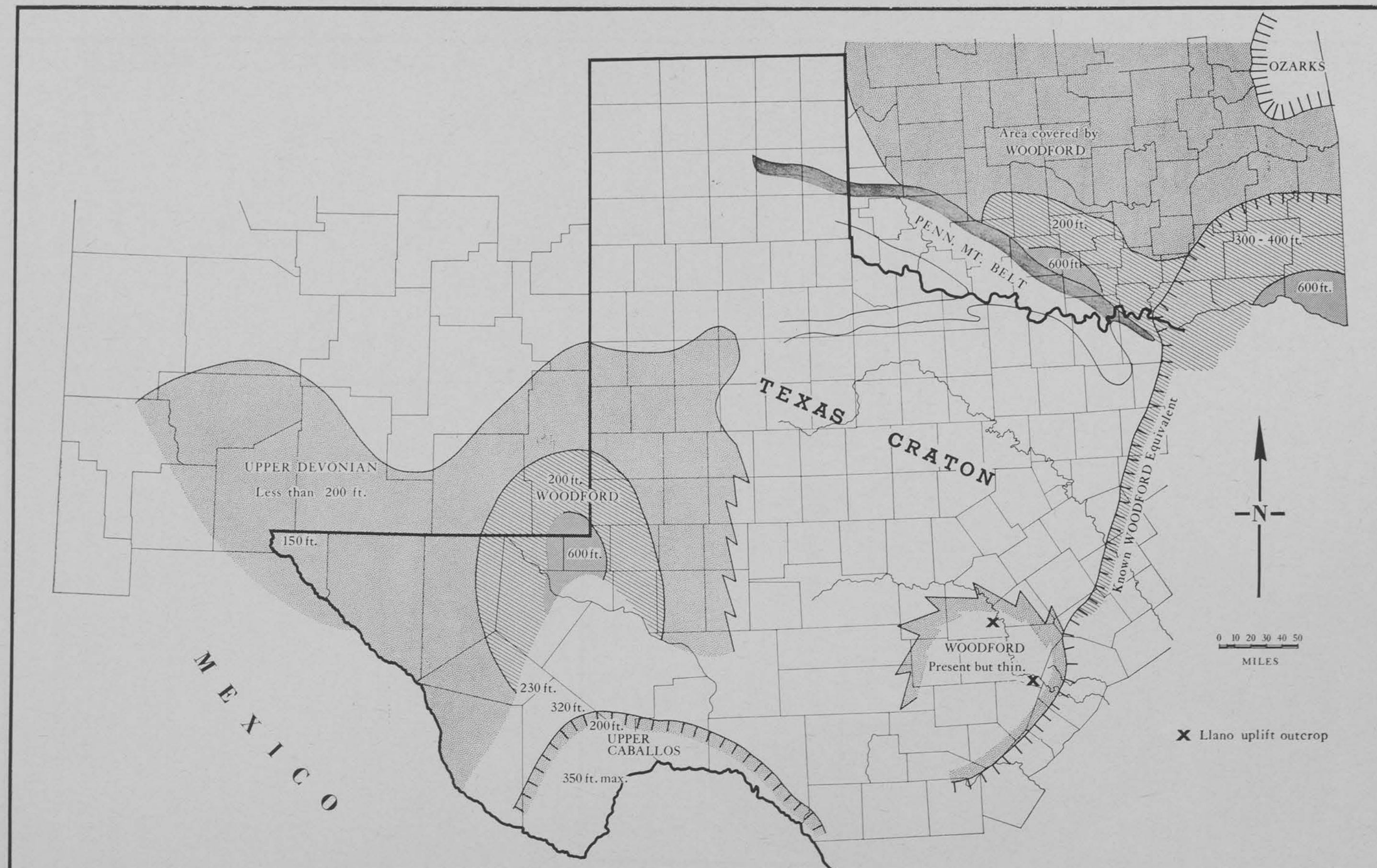


FIG. 51. Interpreted surface and subsurface distribution of Late Devonian (Woodford) strata in the Southwest.

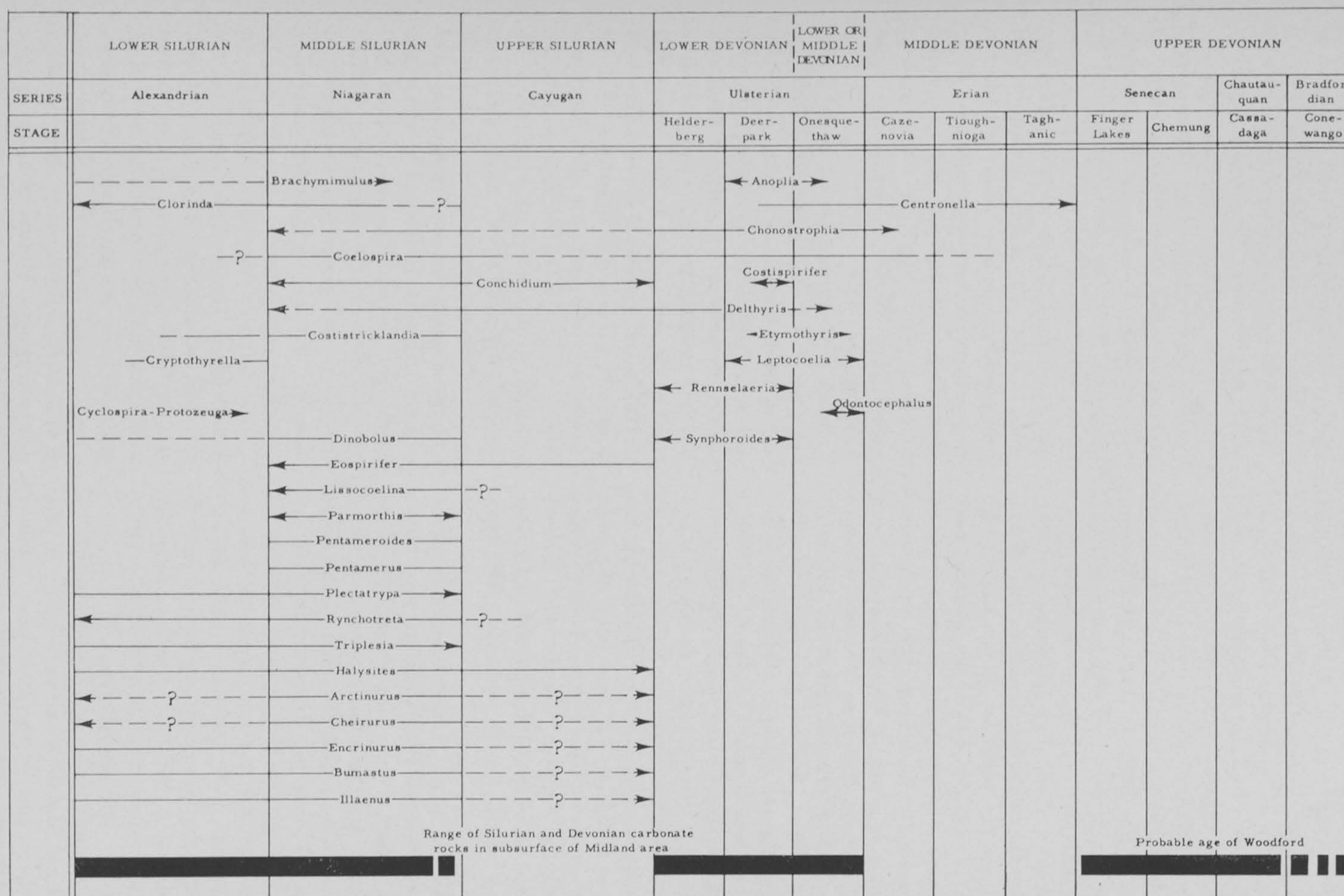


FIG. 52. Range chart of important Devonian and Silurian genera recovered from cores in west Texas. (Mainly after Cooper et al. (1942), Delo (1940), and Shimer and Shrock (1944).)

strata. This may be proved for as much as 300 feet of strata in Andrews, Midland, and Ector counties. The fauna includes the brachiopods *Etymothyris*, *Centronella*, *Leptocoelia*, and *Anoplia* and the phacopid trilobite *Synphoroides* and may range from Oriskany (Deerpark) to Schoharie (Onesquethaw) age. These strata encompass in age both the Frisco of Oklahoma and the Stribling of the Llano uplift. A distinctly Oriskany fauna with *Costispirifer* is known from thin Devonian bioclastic limestone immediately under the Woodford as far north as Gaines County. Ellison's (1950, p. 14) Upper Devonian age assigment for the 90 feet of carbonate rock beneath the Woodford shale in Andrews County is based on linguloid brachiopods and does not appear to be compatible with the ages determined from the fossils reported here. The lower dark cherty member present in the southern Midland basin lies below these faunas and may be of Helderbergian age, but no fossils are known to prove this. Helderbergian fossils a few feet below the Woodford in a well in Andrews County are reported by Stainbrook (*in* Jones, 1953, p. 14) and have also been recovered by the writers from another southern Andrews County well beneath strata correlated lithologically with beds in a nearby well containing Deerpark fossils.

In the west Texas basin, Devonian strata beneath the Woodford and above the Silurian silty limestone (discussed below) are therefore known to range through the Lower Devonian to the lower Middle Devonian. These strata have not yet been given a formal name; it is not the intent of this paper to do so but only to point out their importance as a distinct unit from the underlying Silurian and to recommend that they be given a name by persons more familiar with their detailed lithology and distribution.

Devonian strata rest disconformably on a thick Middle Silurian section. In the southern part of the west Texas basin a lithic boundary between the Devonian and Silurian is recognizable at a change from the dark cherty carbonates of the Devonian downward through a thin dark shale (Devonian) to dark gray and gray-green shales (Silurian) or, more commonly, silty and argillaceous limestones. Decker (1952) recognized Ludlovian (Henryhouse) graptolites, the *Monograptus vomerinus* fauna, from Silurian shales in wells in Crane County, and *Monograptus* fragments have also been recognized by the writers from a well in Upton County. More abundant faunas including brachiopods, corals, ostracodes, trilobites, and pelecypods are known from the Silurian carbonate rocks in the northern part of the west Texas basin, and these indicate Niagaran age (fig. 53). No zonation within this Middle Silurian has been possible.

Below the Silurian clastics in the southern part of the west Texas basin occurs a well-defined carbonate unit, generally called by petroleum geologists the subsurface Fusselman. In parts of the basin this unit contains the same lithic subdivisions in the same order as the Chimneyhill of Oklahoma, namely, a lower oolitic bed, a middle glauconitic limestone, and an upper pink crinoidal unit (Lexicon Committee, 1958, p. 53). Our faunas from the subsurface Fusselman are insufficient to demonstrate conclusively either Niagaran or Alexandrian age. Two wells in Upton County contain small brachiopods and proetid trilobites immediately above the basal oolitic strata. The brachiopods suggest correlation with the Lower Silurian Girardeau and Edgewood formations of Missouri, but more and better material is needed to verify this.

The outcrop Fusselman formation in Trans-Pecos Texas (type area in the Franklin Mountains) is a dolomite unit almost 1,000 feet thick. It has yielded fossils adequate for correlation only in its topmost beds in the Hueco Mountains. They include *Wilsonella*, *Stegerhyncus*, *Whitfieldella*, *Calymene*, and *Illaenus*, demonstrating a Middle Silurian age for these beds. Sections near the base of the Fusselman formation in the Franklin and

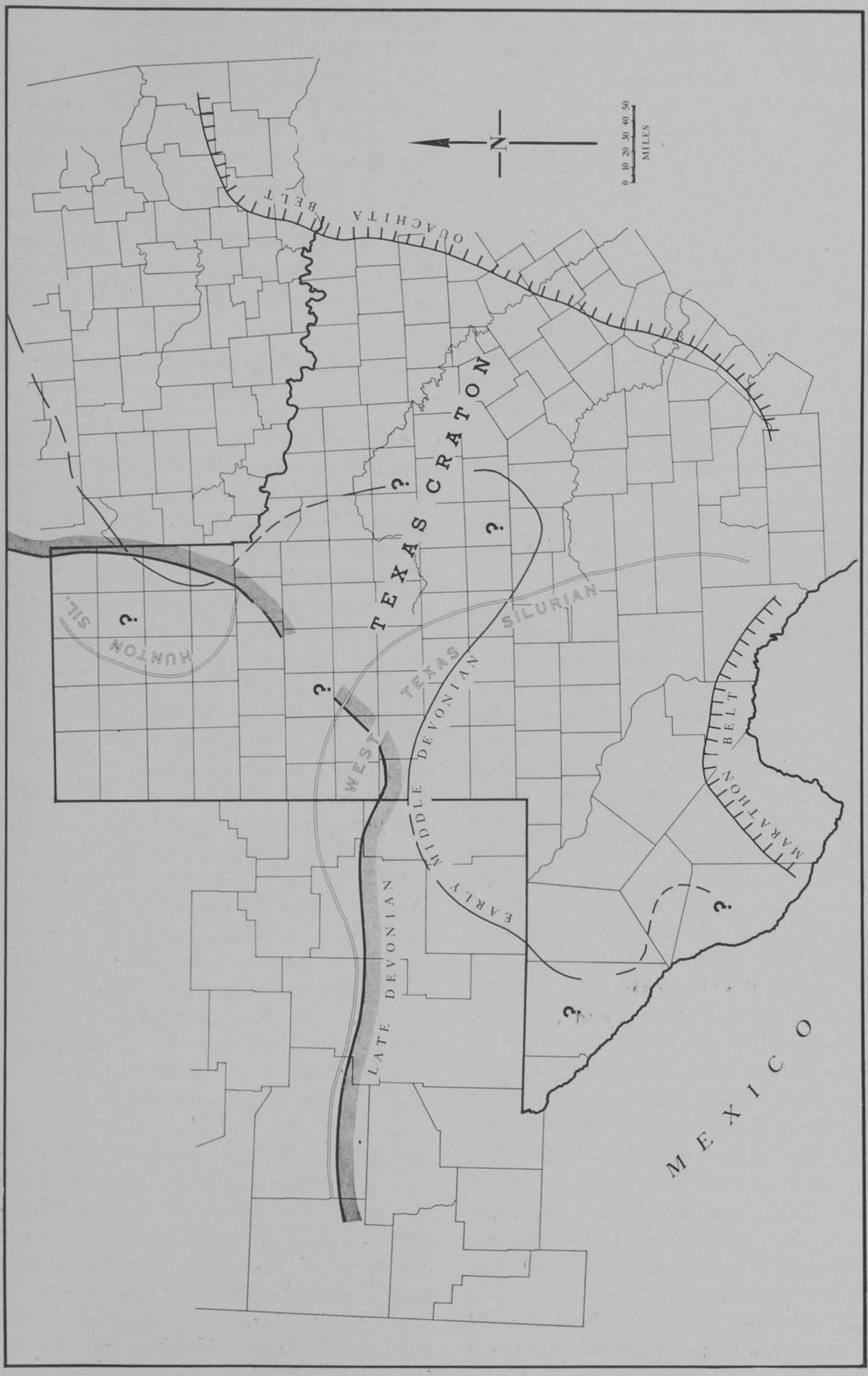


Fig. 53. Conjectured depositional edges of Silurian and older Devonian strata in the Southwest.

Hueco Mountains contain large costate pentamerid brachiopods which indicate a questionable Early Silurian age for the strata, and Pray (1953, pp. 1913–1916) reported an Early Silurian age for the basal remnant of the deeply eroded Fusselman in the Sacramento Mountains.

### New Mexico and West Texas Area of Upper Devonian Outcrops

Overlying the Silurian in Trans-Pecos Texas and in New Mexico is an extensive Late Devonian faunal sequence. None of the west Texas typical carbonate Devonian of Onondagan and older age is present in these western outcrop areas. Instead, dark argillaceous sections with some siltstones and nodular limestone are present. The faunal relations of these strata have been worked out over a period of some years by Stevenson (1945), Cooper (1954), Stainbrook (1947, 1948), Miller and Collinson (1951), and Flower (1958). The faunas range from very latest Devonian down through the Chemung or Finger Lakes stage (Frasnian). Enough is already known of these faunas to assure their correlation with widely scattered units of the western Devonian faunal province, such as the Independence shale and Cedar Valley limestone of Iowa, and the extensive Canadian Rocky Mountain Devonian section. An important correlation between conodont and brachiopod faunas is possible between the Upper Devonian Sly Gap formation of New Mexico and the Independence shale; the former contains numerous Independence brachiopods (Stainbrook, 1935, 1948). The conodonts from the Independence shale, described by Müller and Müller (1947, p. 1069), are found in Hass' conodont zone II. In addition, Müller has identified a key conodont species of Hass' zone I (*Polygnathus linguiformis*) in the middle of the type Canutillo formation just above the lower cherty member. Conodonts from the top beds of the lower cherty member are lower Upper Devonian, according to S. P. Ellison (personal communication) equivalent to Rhinestreet and Geneseo faunas. The black shale in the upper part of the Canutillo is at present considered the Ready Pay equivalent, a western and southern facies of the calcareous Sly Gap shale. The Ready Pay 7 miles west of Hillsboro on New Mexico Highway 180 contains a few brachiopods of Frasnian age. It thus appears that the total New Mexico and Trans-Pecos Texas Devonian section is equivalent to the Woodford of the west Texas basin.

### Ouachita-Marathon Folded Belt Deposits

Although they are separated by many hundreds of miles along the sinuous strike of the Ouachita-Marathon foldbelt, the Caballos formation of west Texas and the Arkansas novaculite of Oklahoma and Arkansas are almost identical units. These units consist generally of cherts and some interbedded varicolored shales and the peculiar siliceous rock, novaculite. The upper parts of both the Caballos and the Arkansas novaculite contain conodonts of Hass' zone III. In addition, the Arkansas novaculite contains practically the complete sequence of Chattanooga shale conodont zones ranging from II in the middle of the middle member up through zones III and VI in the upper part of the middle member (Hass, 1956b, p. 28). Thus, on the basis of conodonts, these units are also fairly well correlated with the Woodford of the Arbuckle Mountains. Graves (1952, p. 610) also has described Woodford conodonts of Hass' zone III (*Palmatolepis perlobata* and *P. subperlobata*) from the middle of the Caballos formation.

In both the Caballos and Arkansas novaculite a lower novaculite member occurs disconformably below the rest of the formation. The lower unit has long been conjectured to be equivalent to the Camden chert of Tennessee (Onesquethawan stage) and to be the only representation in the Ouachita-Marathon geosyncline of the widespread lower Middle and Lower Devonian unit of the basins lying to the north of the geosyncline.

Recently Hass (1956b, p. 27) reported that G. A. Cooper reviewed an identification by Schuchert (Honess, 1923, p. 117) based on a specimen of supposed *Leptocoelia flabellites*, a Camden fossil identified from the top of the lower novaculite of southeast Oklahoma (Pinetop chert), and concluded that its preservation did not warrant an accurate identification. The correlation of the lower novaculite and lower Caballos units with the Onesquethawan stage still rests mainly on their lithic similarity to the Camden and their unconformable stratigraphic contact with Woodford or Woodford equivalent beds above. In the Marathon region the disconformable nature of the lower Caballos unit was first recognized by King (1937, p. 52). Berry and Nielson (1958) have recently demonstrated that it is more restricted in its distribution than the upper Caballos and is overlapped by the upper member. Another possible correlation is noted here (fig. 49): The lower Caballos unit could actually be equivalent to the lower part of the Canutillo chert of the Trans-Pecos Texas outcrops and the lower cherty member of the Woodford identified by Ellison.

In the Marathon foldbelt there is no known Silurian; the Caballos lower novaculite member rests on uppermost Ordovician graptolite-bearing siliceous and argillaceous beds of the Maravillas formation. However, in the Ouachita Mountains a more complex situation exists. The lower part of the Arkansas novaculite (Pinetop chert) in Oklahoma and in large sections of Arkansas apparently rests conformably on a reddish shale unit long known as the Missouri Mountain formation. Hendricks et al. (1947) state that this boundary is gradational. A general statement also indicates that fossil fragments from the Missouri Mountain in southeastern Oklahoma are Silurian in age, but the faunal evidence given is far from conclusive. The Missouri Mountain was equated many years ago with the reddish beds in the Middle Silurian of western Tennessee (Miser and Purdue, 1929, p. 49), and the Silurian age of the formation has become established in the literature generally from one writer's repeating another. However, the Missouri Mountain rests with distinct unconformity on the Blaylock sandstone and overlaps the Blaylock northward to rest on Polk Creek shale (Ordovician). The Blaylock is a massive unit, 1,000 feet thick, of provable Silurian age in the southernmost interior part of the Ouachita geosyncline. Somewhere within the Blaylock, its exact position apparently not known, a Lower Silurian graptolite fauna was discovered many years ago (Miser and Purdue, 1929, p. 45). The Blaylock is probably both Lower and Middle Silurian in age, and the unconformable Missouri Mountain unit above it is probably equivalent to the Helderbergian Haragan shale unit of the nearby Arbuckle Mountains. Proof of this relationship is not now possible through faunal evidence but only through regional stratigraphic consideration. Red beds at this middle Paleozoic interval are certainly not restricted to the Silurian portion of the Hunton in the Arbuckle Mountains (Amsden, 1957, p. 26). The reddish Missouri Mountain argillites may be traced in the subsurface southward along the Ouachita belt to Bell County, Texas, between the Devonian novaculite and Maravillas black graptolitic strata.

Four major stratigraphic sequences in the Silurian and Devonian of the Southwest are demonstrated in figure 49.

*Alexandrian and lower Niagaran.*—The lowest of the important biostratigraphic horizons is of Alexandrian and lower Niagaran age. This includes the Lower Silurian recognized by Pray (1953) in southern New Mexico, the Fusselman limestone of the west Texas subsurface, the Chimneyhill units of the Arbuckles, and the St. Clair-Brassfield-Edgewood-Girardeau of the Ozarks. These thin, pure carbonate deposits apparently represent a widespread Early Silurian sea and are preserved even in some persistently positive areas such as the Ozark dome and the northwestern rim of the west Texas basin. The lithic continuity of the "pink crinoidal" or lower Niagaran portion of this sequence is amazing in view of its thinness. A disconformity separates the Alexandrian from the Niagaran portion of this unit in the Arbuckles, but the Lower-Middle Silurian beds are probably gradational in west Texas.

*Niagaran.*—The Niagaran is the next major stratigraphic unit and, unlike the thin uniform Lower Silurian carbonate strata, includes a variety of well-differentiated facies ranging from fossiliferous marls of the Henryhouse-Brownsport formations to dark graptolitic shales—the "Middle Silurian clastics" of west Texas subsurface—to the thick Niagaran dolomites of the type Fusselman of Trans-Pecos Texas and southern New Mexico. The Niagaran formations are generally thicker than those of Alexandrian age and approach 1,500 feet in the subsurface of New Mexico. It is interesting that in general fossils from the Niagaran beds of the west Texas basin are unlike those of the Henryhouse and Chimneyhill formations of the Hunton group in the Arbuckle Mountains. Only two species of *Leptaena* and one of *Coelospira* (both long-ranging genera) are known in common with the Henryhouse fauna. None of the fossils from the writers' collection at the top of the Fusselman in the Hueco Mountains are like those from Oklahoma. Of all the fossils recovered from the Fusselman cores, those that also occur in the Chimneyhill are species of the trilobite genera *Bumastus* and *Illaenus*, a single species of bryozoan, and a species of the brachiopod *Triplesia*. Better collections from west Texas are needed to indicate whether or not the Alexandrian faunas are different from those of Oklahoma. It is an open question whether any direct connection existed between the west Texas Niagaran sea and the Middle Silurian waters of the Mid-Continent, either through the Ouachita-Marathon geosyncline or across the Texas craton. Different Middle Silurian faunas in the two areas may be explained by deposition in two separated basins. An alternative possibility is that the different lithofacies of the higher Niagaran in the two areas account for the faunal difference.

An unconformity above the Niagaran exists in the Southwest at the position occupied by the evaporites found in basins farther north (New York, Michigan, Alberta). In place of the evaporites there is evidence of truncation of the Silurian beneath the Lower Devonian in the Arbuckles (Amsden, *in* Amsden and Boucot, 1958, p. 16). In west Texas there is a distinct shift in the depositional centers as well as in the post-depositional downwarping areas (fig. 50) between Silurian and Lower Devonian time. This unconformity must represent a considerable period of tectonic activity for it is present in some negative areas as well as in cratonic localities. Thus, subsurface well-to-well correlations suggest that the Hunton group of the Anadarko basin contains the Siluro-Devonian unconformity (Wheeler, 1947) just as does the Arbuckle Mountain section. Log correlation in the west Texas basin across Midland County and northward shows an unconformity between the Middle Silurian clastics and the overlying cherty beds generally assigned to the Devonian. Despite this stratigraphic evidence, the age relation of the unconformity is hard to evaluate, since the Helderbergian fossils found by the writers and those reported by Stainbrook (*in* Jones, 1953) are from Andrews County, where the Siluro-Devonian lithologic contact is harder to determine. The time represented by this unconformity may be slight. There is actually not a very good standard faunal sequence upon which to evaluate the unconformity. Shelly faunas of post-Niagaran Silurian age are not very well known in North America, and the graptolite faunas of the upper Niagaran of Oklahoma and west Texas show identity with the lower Upper Silurian (Ludlovian) graptolites of Britain.

Above the Siluro-Devonian unconformity there appear widely distributed but thin and rather sporadic deposits of Helderberg (Lower Devonian) age which do not form a very well-defined stratigraphic unit (Linden group, Haragan-Bois d'Arc, Pillar Bluff, and Lower Devonian shaly beds of the west Texas basin). The Missouri Mountain shale-slate formation of the Ouachita geosyncline may well be a part of this unit.

*Oriskany to Onondagan.*—Overlying the Helderbergian is an extensive carbonate unit, light colored and highly cherty, ranging in age from Oriskany to Onondagan. Biostratigraphically, the middle portion of this unit consists of Onesquethawan beds, represented in the Camden of Tennessee and the Stribling of the Llano uplift by the same fauna and by the same unusual lithology of novaculite and white chert. The Camden and Stribling are possibly equivalent to the lower novaculite of the Arkansas and Caballos formations in the Ouachita-Marathon belt. Beds of post-Helderbergian through Onondagan ages are generally thin in eastern areas and only remnants occur in the Llano uplift, but they are thickest in west Texas where about 1,100 feet of Lower and lower Middle Devonian has been measured in the southern Midland basin. The top of this sequence in west Texas is probably Onondagan, and beds this young are known with certainty in the Llano uplift. Probably post-Oriskany parts of this sequence have been eroded from the Arbuckles.

There is also an important regional unconformity between the sequence just described and the base of the Woodford or its equivalent. Most, if not all, of the Erian is absent in the Southwest. This time includes the Hamilton group of the New York section and an extensive body of sedimentary rocks of Givetian age (*Stringocephalus* beds) in the Williston and Alberta basins. Although the general areas of deposition of the Woodford are about the same as those of the older Devonian, pre-Woodford truncation is known to have occurred both in west Texas and in Oklahoma, and the areas of thickest preserved Woodford strata are slightly different from those of pre-Woodford Devonian in both basins.

*Woodford-Late Devonian.*—The overlying Woodford-Late Devonian sequence is part of the widespread blanket black shales of the Mid-Continent and northwestern states and represents the last major segment of strata considered herein. The unit has a maximum thickness of about 600 feet, in both west Texas and the Anadarko-Arbuckle region of Oklahoma. Across Texas from southeast to northwest it is represented by several facies: (1) siliceous varicolored shale, chert, and novaculite in the Ouachita-Marathon belt, containing only conodonts and petrified wood; (2) conodont- and spore-bearing black shale in the Woodford of the west Texas and Anadarko basins; and (3) calcareous shales and nodular limestones interbedded with dark shales farther west in New Mexico strata.

# CONJECTURES ABOUT MIDDLE PALEOZOIC TECTONISM AND PALEOGEOGRAPHY

Even though generalized, the subcrop, outcrop, and isopach patterns plotted on figures 50, 51, and 53 make possible several tectonic interpretations, advanced here for further consideration.

(1) The distribution of the eroded remnants of the middle Paleozoic deposits (figs. 50 and 51) outlines very clearly the area of the Texas arch of Adams (1955, p. 238) [equals Concho arch of M. G. Cheney (Galley, 1958, pp. 400, 401)]. This has been described by Flawn (1953, p. 900) as an ancient Precambrian feature, termed the Texas craton. The absence of middle Paleozoic beds in the Fort Worth basin and their restricted occurrence in the Kerr basin suggest that the Anadarko and west Texas basins surrounding the Texas craton are much older features than the Midland, Kerr (?), and Fort Worth basins, which did not begin to subside strongly until late Paleozoic orogeny commenced. The numerous thin remnants of pre-Woodford carbonates scattered over such great distances across the Llano uplift could only have been preserved from a section made up of several rather thin formations of varying geologic ages and separated by numerous disconformities. The fact that Late Devonian and Mississippian rocks are also preserved in the Llano uplift sink holes must mean that the Llano area stood as a tectonically positive element for the greater portion of middle Paleozoic time. The writers conjecture that thin carbonate strata of middle Paleozoic age were probably deposited in both the Kerr and Fort Worth basins (as on the Llano uplift).

(2) The above generalization may not be correct for the Kerr basin. A 200-foot thick limestone of post-Ellenburger age has recently been found by V. E. Barnes in Kerr County (1959, Pl. 1). While conceivably this may be older Paleozoic (Simpson?), lithologically it most resembles the Devonian of the Llano uplift. Coupled with the wide distribution of Devonian remnants in the Llano uplift, this may indicate that farther south a yet thicker Devonian section exists in the Kerr basin. So far as the writers know, no other evidence exists of middle Paleozoic deposition in the Kerr basin.

(3) An interesting restriction of the Silurian is observed. This may be explained either by extensive pre-Devonian erosion or by a more limited area of deposition for the Silurian (or both). In the Ouachita geosyncline the Silurian is restricted to the thick Blaylock sandstone of the interior orogenic belt (southeastern Ouachitas), if one accepts the Missouri Mountain formation as being of Devonian age. This is in accord with the absence of the Silurian under Lower Devonian in the extreme eastern Arbuckle Mountains and places the area of no Silurian in the Arbuckles next to a wide area of no Silurian in the north and northwestern part of the Ouachitas. Apparently the Silurian is widespread in the Anadarko basin and is nowhere overlapped by the Devonian. Neither in west Texas is there overlapping by the older Devonian on the Silurian *along the presently preserved limits of the Silurian*. However, the map (fig. 50) suggests that on the east side of the west Texas basin overlapping did occur but pre-Woodford erosion stripped back the older Devonian, leaving only a few patches such as those found in the Llano uplift. Thus, the Silurian is presumed to be missing over the Texas craton and along the buried Ouachita trough at least as far south as Williamson County—the southern limit of recognizable Ouachita pre-Mississippian sediments. The Marathon section is very similar to the Ouachita section, and no Silurian is to be expected in the presently exposed foldbelt of the Marathon Basin which is comparable in facies position to the outermost (northwestern) part of the Ouachita foldbelt. If this geographic re-

striction of the Silurian truly reflects its depositional areas, an explanation is offered for faunal differences already cited between the west Texas and Hunton Silurian strata.

(4) There is evidence of considerable truncation in pre-Devonian time. Some areas in which Early Devonian overlaps the Silurian have been mentioned both in the Arbuckles and in west Texas. Log correlations from the southern Midland basin northward show that the Devonian onlaps itself to some degree in this direction. Barnes et al. (1947, p. 140) have already pointed out that on the Texas craton, Middle Devonian deposits rest on Ellenburger 1,000 feet lower in the section on the western side of the Llano uplift than on the eastern side. This erosion occurred at some time between medial Ordovician and Early Devonian. From what is known of regional unconformities in west Texas, erosion probably occurred during Silurian or between Silurian and Devonian periods. Additional evidence of tectonic activity toward or at the end of Silurian time is the shift of the axes of depositional and/or structural basins in west Texas, seen when comparing isopachs of the Silurian, Lower-Middle Devonian, and Upper Devonian strata.

(5) Evidence of the Siluro-Devonian unconformity is partially obscured by subsequent erosion periods. Pre-Woodford truncation of the older Devonian removed much of the latter from the edges of the west Texas and Anadarko basins before the Woodford deposits were laid down. Woodford strata widely overlap the Early-Middle Devonian in the west; the Upper Devonian is widespread over western New Mexico and Arizona, resting on Silurian in much of this area. Much greater uplift and truncation took place in Late Mississippian and Early Pennsylvanian time in the Southwest, for the Woodford subcrop pattern around almost all of the west Texas basin and along the Pennsylvanian mountain belt of the Wichita-Amarillo axis coincides closely with the subcrop of the underlying Silurian.

# REFERENCES


Adams, J. E. (1955) Mid-Paleozoic paleogeography of central Texas: Shale Shaker Digest, Semco Color Press, Oklahoma City, Oklahoma, pp. 238–241.

Amsden, T. W. (1951) Brachiopods of the Henryhouse formation (Silurian) of Oklahoma: Jour. Paleont., vol. 25, pp. 69–96.

——— (1957) Stratigraphy of the Hunton group: Oklahoma Geol. Survey, Circ. 44, 57 pp.

———, and Boucot, A. J. (1958) Stratigraphy and paleontology of the Hunton group in the Arbuckle Mountain region: Oklahoma Geol. Survey, Bull. 78, 199 pp.

Barnes, V. E., et al. (1959) Stratigraphy of the pre-Simpson Paleozoic subsurface rocks of Texas and southeast New Mexico: Univ. Texas Pub. 5924, 2 vols., 837 pp.

———, Cloud, P. E., Jr., and Warren, L. E. (1947) Devonian rocks of central Texas: Bull. Geol. Soc. Amer., vol. 58, pp. 125–140.

Berry, W. B. N., and Nielson, H. M. (1958) Revision of Caballos novaculite in Marathon region, Texas: Bull. Amer. Assoc. Petrol. Geol., vol. 42, pp. 2254–2259.

Cloud, P. E., Jr., Barnes, V. E., and Hass, W. H. (1957) Devonian-Mississippian transition in central Texas: Bull. Geol. Soc. Amer., vol. 68, pp. 807–816.

Cooper, G. A. (1954) Unusual Devonian brachiopods: Jour. Paleont., vol. 28, pp. 325–332.

——— et al. (1942) Correlation of the Devonian sedimentary formations of North America: Bull. Geol. Soc. Amer., vol. 53, pp. 1729–1794.

Decker, C. E. (1935) Graptolites from the Silurian of Oklahoma: Jour. Paleont., vol. 9, pp. 434–446.

——— (1936) Some tentative correlations on the basis of graptolites of Oklahoma and Arkansas: Bull. Amer. Assoc. Petrol. Geol., vol. 20, pp. 301–311.

——— (1952) Texas graptolites change supposed Devonian zone to Silurian: Bull. Amer. Assoc. Petrol. Geol., vol. 36, pp. 1639–1641.

Delo, D. M. (1940) Phacopid trilobites of North America: Geol. Soc. Amer., Special Paper No. 29, 135 pp.

Dunbar, C. O. (1919) Stratigraphy and correlation of the Devonian of western Tennessee: Tennessee State Geol. Survey, Bull. 21, 127 pp.

Ellison, S. P. (1950) Subsurface Woodford black shale, west Texas and southeast New Mexico: Univ. Texas, Bur. Econ. Geol. Rept. Invest. No. 7, 20 pp.

———, and Wynn, W. T. (1950) Devonian microfossils, Andrews County, Texas: Amer. Jour. Sci., vol. 248, pp. 794–799.

Flawn, P. T. (1953) Texas basement rocks, a progress report: Bull. Amer. Assoc. Petrol. Geol., vol. 37, p. 900.

Flower, R. H. (1958) Cambrian to Mississippian beds of southern New Mexico: Guidebook of the Hatchet Mountains and the Cooks Range—Florida Mountain areas, Roswell Geol. Soc., pp. 61–78.

Galley, J. E. (1958) Oil and geology in the Permian basin of Texas and New Mexico, *in* Habitat of oil, Amer. Assoc. Petrol. Geol., pp. 395–446.

Graves, R. W. (1952) Devonian conodonts from the Caballos novaculite: Jour. Paleont., vol. 26, pp. 610–612.

Hass, W. H. (1956a) Age and correlation of the Chattanooga shale and the Maury formation: U.S. Geol. Survey Prof. Paper 286, 38 pp.

——— (1956b) Conodonts from the Arkansas novaculite, Stanley shale, and Jackfork sandstone: Guidebook of Ouachita Field Conference, southeastern Oklahoma, Ardmore Geol. Soc., pp. 27–33.

Hendricks, T. A., Gardner, L. S., Knechtel, M. M. (1947) The geology of the western part of the Ouachita Mountains in Oklahoma: U.S. Geol. Survey Oil and Gas Invest., Prelim. Map 66, sheet 1.

Honess, C. W. (1923) Geology of the southern Ouachita Mountains of Oklahoma: Oklahoma Geol. Survey, Bull. 32, pt. 1, 278 pp.

Jones, T. C. (1953) Stratigraphy of the Permian Basin of west Texas. . . . ., West Texas Geol. Soc., pp. 12–18.

King, P. B. (1937) The geology of the Marathon region: U.S. Geol. Survey Prof. Paper 187, 148 pp.

Lexicon Committee of West Texas Geological Society (1958) Lexicon of pre-Pennsylvanian stratigraphic names of west Texas and southeastern New Mexico, ix, 153 pp.

Miller, A. K., and Collinson, Charles (1951) A clymenoid ammonoid from New Mexico: Amer. Jour. Sci., vol. 249, pp. 600–603.

Miser, H. D., and Purdue, A. H. (1929) Geology of the DeQueen and Caddo Gap quadrangles, Arkansas: U.S. Geol. Survey Bull. 808, 195 pp.

Müller, K. J., and Müller, E. M. (1957) Early Upper Devonian (Independence) conodonts from Iowa, Part 1: Jour. Paleont., vol. 31, pp. 1069–1108.

Pray, L. C. (1953) Upper Ordovician and Silurian stratigraphy of Sacramento Mountains, Otero County, New Mexico: Bull. Amer. Assoc. Petrol. Geol., vol. 37, pp. 1894–1918.

Shimer, H. W., and Shrock, R. R. (1944) Index fossils of North America, John Wiley & Sons, Inc., New York, 837 pp.

Stainbrook, M. A. (1935) A Devonian fauna from the Sacramento Mountains near Alamogordo, New Mexico: Jour. Paleont., vol. 9, pp. 709–714.

——— (1947) Brachiopoda of the Percha shale of New Mexico and Arizona: Jour. Paleont., vol. 21, pp. 297–328.

——— (1948) Age and correlation of the Devonian Sly Gap beds near Alamogordo, New Mexico: Amer. Jour. Sci., vol. 246, pp. 756–790.

Stevenson, F. V. (1945) The Devonian of New Mexico: Jour. Geol., vol. 53, pp. 217–245.
and oil possibilities: World Oil, vol. 127, nos.

Wheeler, R. R. (1947) Anadarko basin geology 4, 5, 7, 9.

# Paleozoic History of the Fort Stockton–Del Rio Region, West Texas

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ABSTRACT

Throughout Paleozoic time to the close of the Mississippian, the Fort Stockton-Del Rio region experienced only mild structural activity; the associated sedimentary rocks are relatively thin and consist chiefly of carbonates except for the dark shale of the Woodford and overlying younger Mississippian sediments. In Early Pennsylvanian time a major geosyncline developed, was filled with thick clastic sediments, and was later compressed into the Marathon folded belt. In early Permian Wolfcamp time the final phase of this late Paleozoic orogeny occurred. The Val Verde geosyncline formed principally during this time and received at least 14,000 feet of clastic sediments. It is north of the older geo-syncline and is the main structural feature of the Fort Stockton-Del Rio region. After Wolfcamp time the region returned to a condition of crustal quiet. This Paleozoic history has been revealed mostly by wells drilled in the past few years for oil and gas, among which is the deepest boring ever sunk into the earth.

## INTRODUCTION

A line drawn from a point several miles north of Del Rio northwest to and beyond Fort Stockton locates roughly the axis of a great structural trough, one of the major subdivisions of the Permian basin of west Texas and southeastern New Mexico. This structural feature is hereafter referred to as the Val Verde geosyncline, and the main purpose of this paper is briefly to outline the Paleozoic history of the region of which this geosyncline is the principal element.

In figure 54 are outlined the major structural elements of west Texas as established in the late Paleozoic. During most of this time the Delaware basin was an integral part of the great trough which extended southeast beyond Del Rio; therefore, throughout this paper the term "Val Verde geosyncline" refers to the whole structure. This geosyncline lies south of the Eastern shelf and south of the south rim of the Midland basin, an east-west arch for which no generally accepted name has appeared. Farther west the Val Verde geosyncline is bounded on the north and northeast by the Central Basin platform, the southern element of which is the Fort Stockton high. All of these subdivisions of the Permian basin are fairly well understood geologically from information furnished by the many wells drilled for oil and gas.

South of the Val Verde geosyncline lies the Marathon folded belt, an entirely different geological province and one poorly understood except in the limited area of the Marathon Basin. In particular, the precise location and nature of the boundary between the folded belt and the geosyncline to the north of it remain at present a fascinating enigma. Northwest of the Marathon Basin the Val Verde geosyncline is flanked on the west by the Diablo platform, an uplift of the foreland like the Central Basin platform on the eastern side of the geosyncline.

Throughout much of the Fort Stockton-Del Rio region strata of Cretaceous age, mostly Lower Cretaceous, occur at the surface, and the thickness of this sequence

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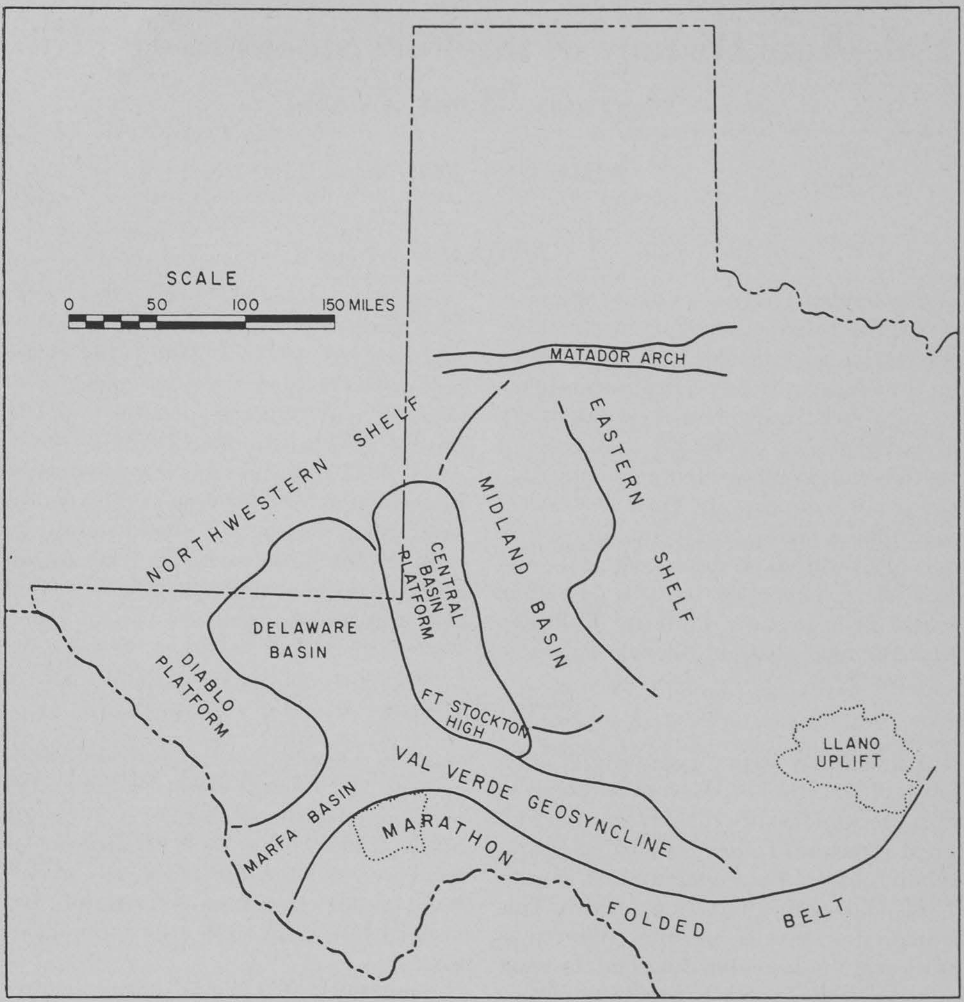



FIG. 54. Major late Paleozoic structural elements of west Texas and southeastern New Mexico.

varies up to 3,000 or more feet. These strata are nearly flat-lying and effectively conceal the character and structure of the underlying Paleozoic formations; for a long time well data were quite limited and inadequate. However, in recent years, following the discovery of important gas reserves in the Val Verde geosyncline, a number of deep tests have been drilled, some of them among the deepest borings ever put down. The information thus brought to light makes possible a preliminary sketch of the geology of the Val Verde geosyncline, but it is to be emphasized that much remains to be uncovered and conclusions reached today are subject to considerable future revision. This summary of Paleozoic history should, therefore, be considered a progress report only.

The Paleozoic record in the Fort Stockton-Del Rio region is complete; at least, each of the seven systems is represented by rock strata, although some of the units are thin. This Paleozoic section is summarized in figure 55 in which the principal rock units are tabulated and the lithology is represented graphically in a very generalized manner. The graphic column is not drawn to scale. Lithologically the entire section falls naturally into three major subdivisions: The lower one includes all strata from the base of the Cambrian to the base of the Woodford formation and consists very largely of carbonates; chert occurs in the upper members. The middle subdivision extends from the base of the Woodford formation to the top of the Leonard of middle Permian age and consists primarily of dark shale but in places includes much limestone. The uppermost and smallest part of the Paleozoic includes only the Permian above the Leonard; here are the siltstones and fine sandstones, the platform dolomites, and the evaporites typical of the Permian basin. Such, in brief, are the major groups of sediments; in the paragraphs that follow the smaller units are discussed in order and in more detail.

## Pre-Woodford History

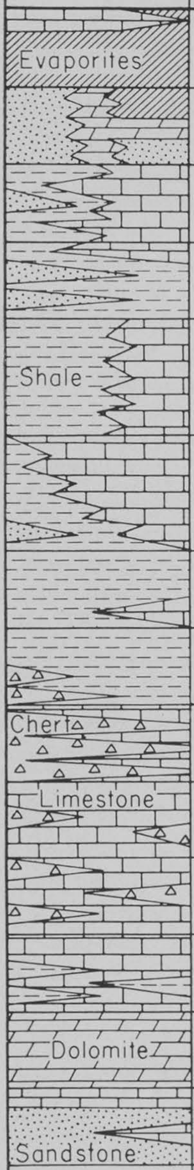
In spite of great depths a few wells within the Val Verde geosyncline have actually reached the Precambrian basement. In the Puckett field of southern Pecos County this basement is a quite ordinary granite, and P. T. Flawn (personal communication) reports that this granite has been determined to have an age of 900 million years. Since the immediately overlying Upper Cambrian beds have an age of something less than 500 million years, it is apparent that a very great span of time is represented by the unconformity at the base of the Paleozoic. There was ample time for the development of the peneplain which is assumed to have existed when the earliest Paleozoic sediments were deposited.

These first Paleozoic deposits were sandstones of Upper Cambrian age. The sands are mostly medium grained, noticeably coarser than typical Permian sands of the same region, and commonly glauconitic. Some limestone is present and at the top there is a layer of limestone or dolomite which closely resembles the overlying Ellenburger, so that in places the boundary is difficult or impossible to identify.

Figure 56 is the first of several maps which depict the present thickness and distribution of the several members of the Paleozoic. All of these maps cover the same geographic area, which is shown in outline on the regional map (fig. 54). Figure 56 is an isopach map of the Cambrian section and reveals a progressive south to north thinning of these sediments. However, the two areas where no Cambrian is found are areas from which the sediments were eroded in late Paleozoic; they are not areas of non-deposition.

These Cambrian thickness variations, as well as the character of the beds and detailed correlations of members within the formation, support the interpretation that the Cambrian sediments are near-shore deposits of a sea which transgressed northward over a well-developed peneplain. In this respect these strata resemble the Trinity section of the Lower Cretaceous of this same region.

The Cambrian sands were followed without interruption by the deposition of dolomites and limestones which belong to the Ellenburger group. Figure 57 presents the present thickness of these strata, and it should be noted that most of the local variations are due to subsequent uplift and erosion. At time of deposition the Ellenburger of this region probably had an approximately uniform thickness of about 1,500 feet and extended far beyond the




| GROUP FORMATION | SERIES | SYSTEM |
|---|---|---|
| | OCHOA | PERMIAN |
| DELAWARE MTN. | GUADALUPE | PERMIAN |
| BONE SPRING | LEONARD | PERMIAN |
| | WOLFCAMP | PERMIAN |
| | CISCO | "UPPER" PENN. |
| | CANYON | "UPPER" PENN. |
| | STRAWN | "UPPER" / "LOWER" PENN. |
| | BEND | "LOWER" PENN. |
| | MORROW, SPRINGER | "LOWER" PENN. |
| | | MISS. |
| WOODFORD | | MISS. / DEVONIAN |
| | | DEVONIAN |
| | | SILURIAN |
| MONTOYA | UPPER | ORDOVICIAN |
| SIMPSON | MIDDLE | ORDOVICIAN |
| ELLENBURGER | LOWER | ORDOVICIAN |
| | | CAMBRIAN |
| PRECAMBRIAN | | |

FIG. 55. Stratigraphic table for Fort Stockton-Del Rio region.


FIG. 56. Present thickness (in feet) of Cambrian strata.

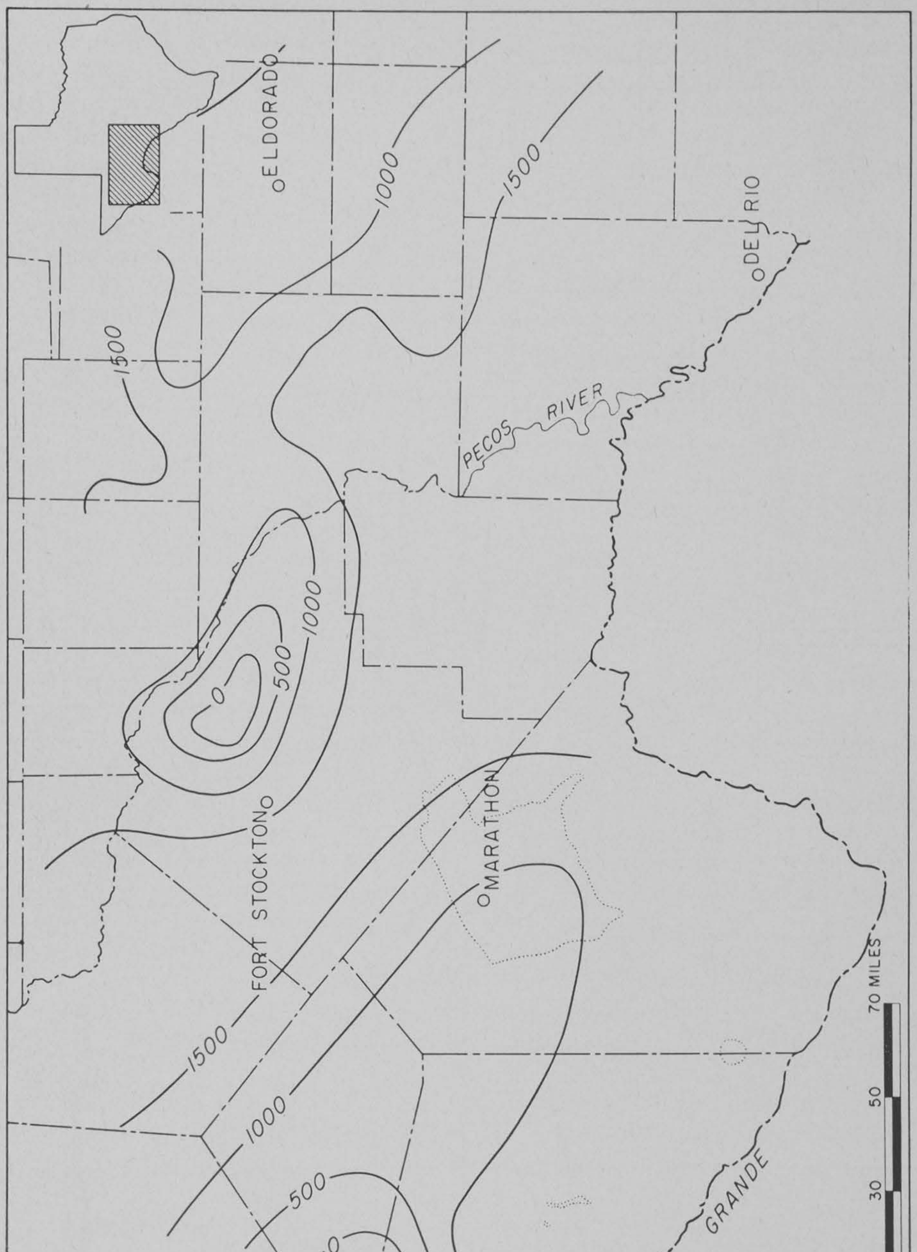



FIG. 57. Present thickness (in feet) of Ellenburger group.

area of figure 57, covering much of Texas and some adjoining states.

Throughout the Fort Stockton-Del Rio region the Ellenburger is a quite uniform deposit of crystalline dolomite with which in some localities are included beds of lithographic limestone. Minor amounts of chert are present in places, and here and there rounded, medium to coarse sand grains "float" in the dolomite. This facies is generally quite similar to the facies of the Ellenburger present over the Central Basin platform and Midland basin; it is distinctly different from the limestone of the Marathon formation which is of equivalent age at its outcrops within the Marathon Basin. A particularly marked contrast is furnished by Slick-Urschel Oil Company No. 1 Mary Decie, a wildcat only a few miles northwest of Marathon limestone outcrops near the town of Marathon. This well was drilled through the Dugout Creek overthrust, a major low-angle thrust described in detail by King (1937), and encountered Ellenburger in a dolomite facies similar to that to the north.

Deposition of the Ellenburger group was brought to a close by a marked contraction of the Ordovician sea. The resulting constricted sea occupied a broadly elliptical area which included the Fort Stockton-Del Rio region and extended north into southeastern New Mexico and the adjoining counties of the Texas south plains. This shallow structural depression has been named the Tobosa basin by Galley (1958). As shown in figure 58, a maximum of 3,000 feet of sediments accumulated in this basin during the time interval from the close of Ellenburger deposition to the beginning of Woodford time in late Devonian. The isopach lines show the original shape of the basin fairly well, but the sharp minor irregularities are the result of subsequent uplift and erosion.

During this specified time interval conditions of sedimentation in the Fort Stockton-Del Rio part of the Tobosa basin remained constant, and with only minor interruptions the basin was filled with a succession of similar deposits, mostly limestones. The lower two-thirds of this limestone sequence belongs to the Simpson group of the Middle Ordovician. In Pecos County, the Simpson attains a thickness of 2,300 feet, its maximum for the entire west Texas area. In the Fort Stockton-Del Rio region this group is composed largely of argillaceous limestone; sand is present in only insignificant amounts, and shale is proportionately less prominent than it is to the north over the Central Basin platform.

Overlying the Simpson group is the Montoya formation, which is somewhat more extensive laterally, particularly on the west flank of the Tobosa basin. The Montoya formation has a maximum thickness of 500 feet and consists predominantly of limestone but, unlike the Simpson, contains significant amounts of chert. The chert is dark brown to bluish gray and smooth to translucent.

Overlying the Montoya is an even thinner unit, a limestone of Silurian age, 0 to 250 feet thick. The lower member of this unit is the Fusselman limestone and in places it is the only part of the Silurian present. The Fusselman is 0 to 200 feet thick and is a light-colored to nearly white limestone which is more obviously crystalline than associated strata. In most localities it contains small amounts of dense white chert.

Conformably above the Fusselman is the uppermost unit of the sequence of lower Paleozoic carbonates of the Tobosa basin. This is a chert and limestone formation of Devonian age, which is 0 to 300 feet thick, and it is common practice in west Texas to refer to this formation as "The Devonian," the Devonian age of the overlying Woodford formation being disregarded. This limestone is somewhat similar to that of the underlying Silurian but is generally darker in color, and in some areas of the Fort Stockton-Del Rio region it grades into a dolomite facies. Chert is more abundant in this unit than in any other portion of the

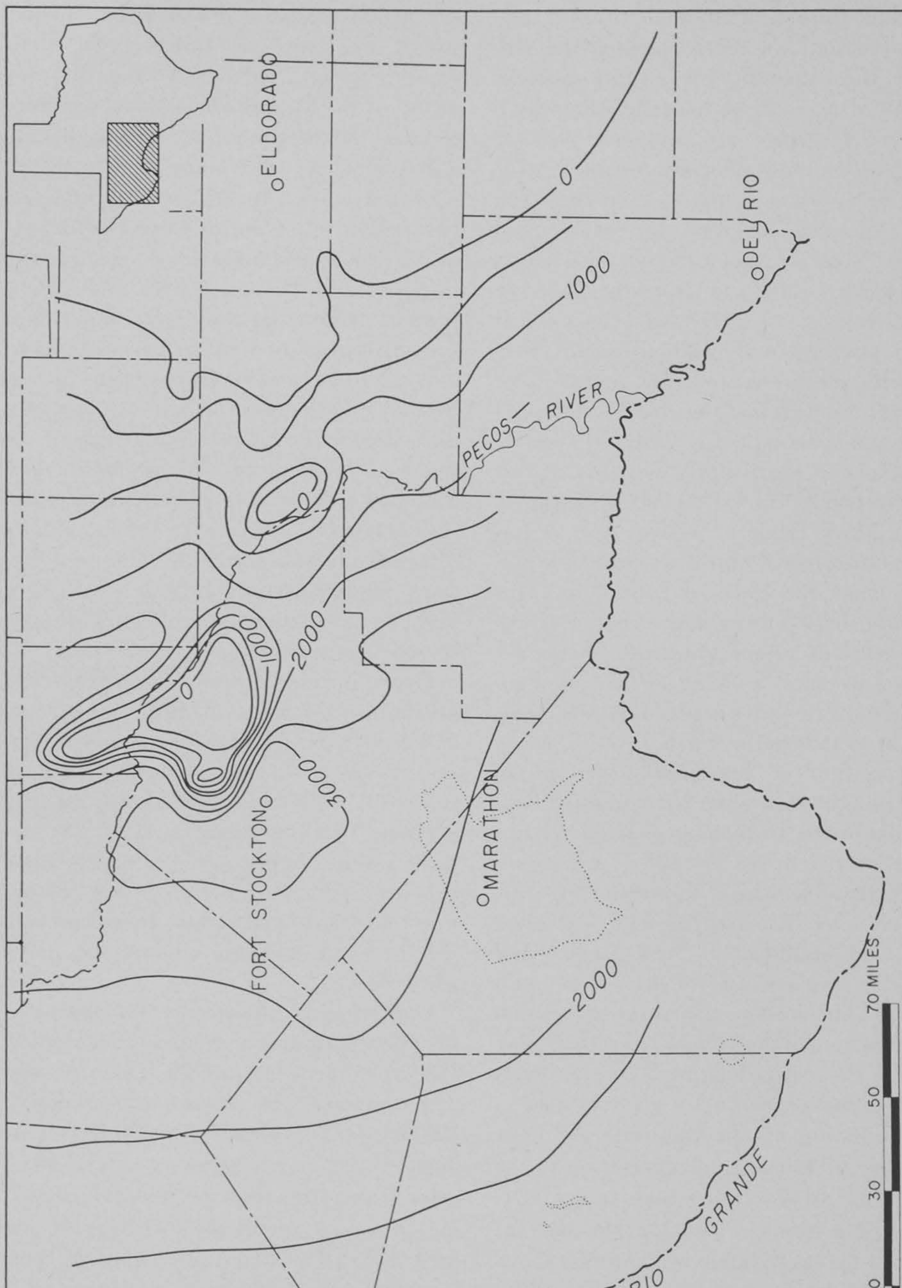



FIG. 58. Present combined thickness (in feet) of strata between top of Ellenburger group and base of Woodford formation.

stratigraphic column, and in and near the Marathon Basin the unit is almost wholly chert. This chert is commonly novaculitic and light colored, although darker shades also occur.

### Woodford to Wolfcamp History

The Woodford formation is the lowermost member of the second and middle sequence of Paleozoic rocks, those composed primarily of dark shale, and it marks the beginning of an important change in the sedimentary history of the region. In figure 59 the combined thickness of the Woodford and the overlying unit of Mississippian age is shown by isopachs. This entire group of sediments consists largely of brownish-black to dark brown shale which is typically softer and contains less sand and silt than the overlying Pennsylvanian shales. The Woodford part of the section contains some dark chert which is most abundant along the southern flank of the Val Verde geosyncline. The section above the Woodford contains in its lower part some argillaceous limestone. In the Fort Stockton-Del Rio region limestone is a relatively minor constituent of the Mississippian, but as the unit is traced north limestone becomes increasingly abundant.

The thickness map shows that these Woodford and overlying younger Mississippian sediments attain a maximum thickness of 2,000 feet at the northwest edge of the map, near the heart of the Tobosa basin. The areas of no sediments appear to represent erosion and not non-deposition. In fact, the Mississippian sea spread far beyond the map area, and it may be said that at this time the Tobosa basin ceased to exist as a distinct structural entity.

All the Paleozoic strata from the first Cambrian beds to the close of the Mississippian indicate that during this long interval the crust in this area experienced only broad mild regional upward and downward movements. This peaceful era was closed at the beginning of Pennsylvanian time when major orogeny caused an uplift near the present Texas-Mexico border and a great Early Pennsylvanian geosyncline developed immediately north of this uplift (King, 1937, p. 135; Hall, 1956) and was filled with clastics eroded from it. These events are reflected in figure 60, the first of two isopach maps of the Pennsylvanian sediments of the Fort Stockton-Del Rio region. In this figure the term "lower" Pennsylvanian is for convenience chosen to mean all strata from the base of the Pennsylvanian up to and including a widespread limestone in which lower Strawn fusulines are commonly found.

The "lower" Pennsylvanian thickness map presents two markedly contrasting areas. To the south is a great linear trough where sediments during this time interval accumulated to a thickness of at least 10,000 feet. This amount is in striking contrast to all preceding periods of the Paleozoic, none of which experienced more than a small fraction of this degree of sedimentation. Unfortunately, information about this trend is entirely inadequate. Only a few deep wells have reached these "lower" Pennsylvanian beds along the northern flank of the trough; the south flank is completely unknown. Practically all of our information is obtained from surface outcrops in the Marathon Basin (King, 1937) where the Tesnus, Dimple, and Haymond formations belong to this part of the Pennsylvanian system. These sediments are predominantly shale and sandstone derived from highlands not far to the south. Here another element of uncertainty must be noted. The present isopachs do not show the original site of deposition of these thick "lower" Pennsylvanian beds. These lines include the effects of northward overthrusting which occurred in early Pennsylvanian and again at the close and resulted in the Marathon folded belt.

In contrast to these thick geosynclinal deposits are the much thinner sediments which extend over a far broader area to the north beyond the geosyncline. This

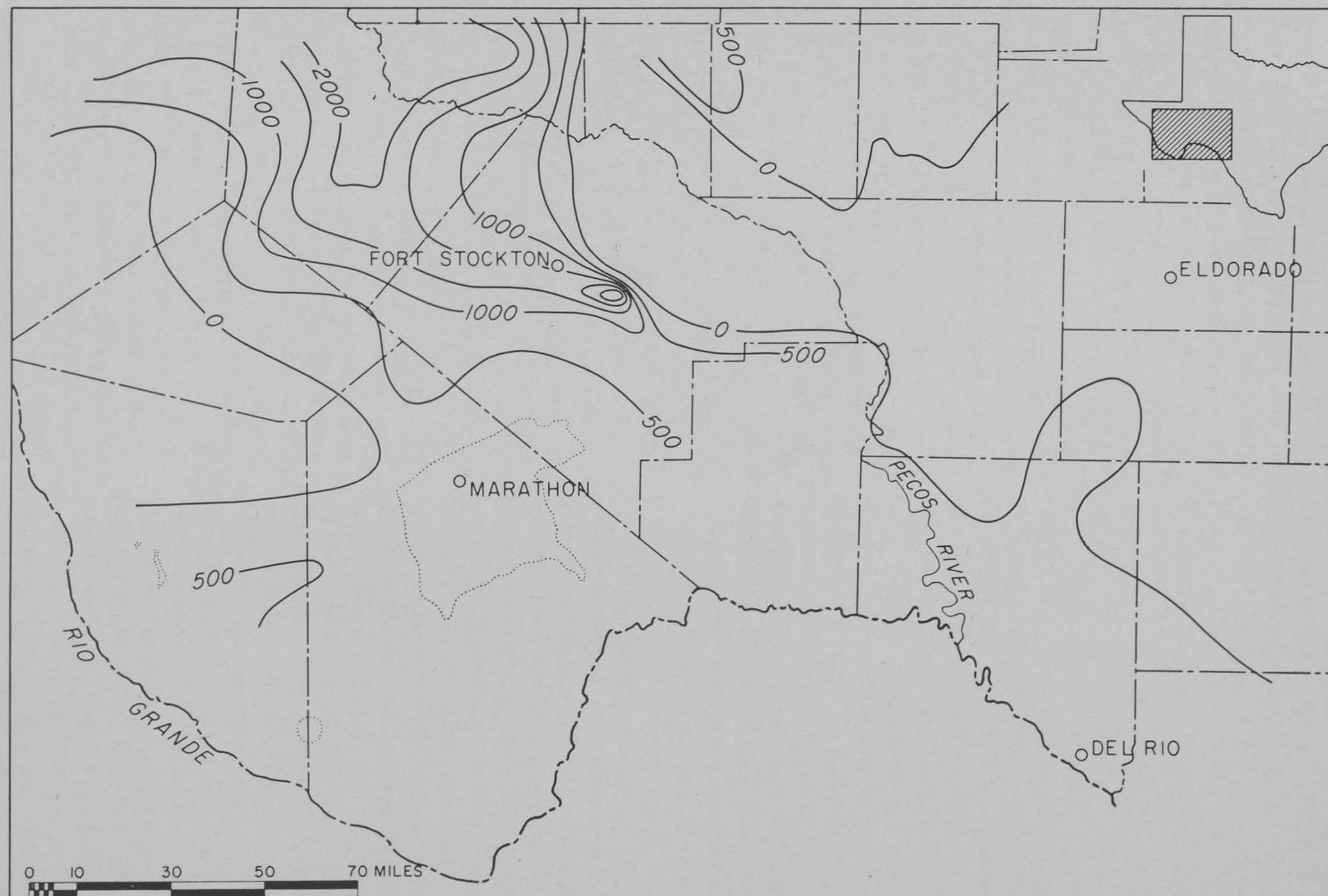



FIG. 59. Present over-all thickness (in feet) of Mississipian and Woodford strata.

unit varies in thickness from 100 to 700 feet and consists predominantly of limestone, which in places is cherty or shaly and in places is fragmental and may be fossiliferous. The upper part of this limestone is lower Strawn in age; the lower part may in places be of Bend age but is nowhere as old as the lower beds in the trough to the south where representatives of Springer and Morrow time are present.

Three areas in which no Pennsylvanian sediments are present appear on the map. The two areas in the northwestern part of the map are over the Diablo and Central Basin platforms, and there the absence of sediments is probably due both to nondeposition and to subsequent erosion. Near Del Rio is a third area where Lower Cretaceous strata rest on a series of incipiently to weakly metamorphosed sedimentary rocks, a considerable part of which are carbonates. No fossils have been found, but probably both Paleozoic and Precambrian rocks are present. It seems likely that "lower" Pennsylvanian deposits also once covered the area, but strong uplift which induced the slight metamorphism was followed by removal of all late Paleozoic strata. The writer suggests that this final erosion took place about the beginning of the Permian.

The "lower" Pennsylvanian map presents the first clear-cut evidence of the beginning of the Val Verde geosyncline. Northwest of Fort Stockton the map reveals a shallow depression between the newly risen Diablo and Central Basin platforms; this low later deepened into the Delaware basin portion of the geosyncline. Southeast of Fort Stockton across Terrell and Val Verde counties the northern fringe of the Pennsylvanian geosyncline partly lies along the trend of the Val Verde geosyncline. This accords with the long-held concept (Cheney, 1929) that geosynclinal formation progressed inward on the continent during late Paleozoic orogeny.

For the purposes of this paper "upper" Pennsylvanian means simply that part of the Pennsylvanian above the lower Strawn limestone. In marked contrast to the "lower" Pennsylvanian section, the upper part is thin, and the implication is clear that it was a time of crustal stability but of shorter duration than the earlier Pennsylvanian.

The thickness map (fig. 61) reveals no area of the "upper" Pennsylvanian much over 2,000 feet thick, and several areas now have no sediments of this age. These blank areas occur over pre-existing highs, and perhaps no sediments were ever deposited. Close to these highs "upper" Pennsylvanian limestone of reef facies accumulated, and in the intervening basins, such as the Midland basin, only very thin deposits of dark shale accumulated contemporaneously. These thin shales are the starved-basin sediments described by Adams et al. (1951). The Delaware basin segment of the Val Verde geosyncline was a starved basin at this time.

Southward in the Marathon Basin "upper" Pennsylvanian is represented by the Gaptank formation. These beds are found only in the northern part of the basin, and King (1937) believes that they were never deposited much south of their present outcrop. They consist mainly of clastics and some interbedded limestone of a near-shore facies, and the whole unit is about 1,800 feet thick (King, 1937, p. 74). The Gaptank formation thus suggests that by the beginning of "upper" Pennsylvanian time the southern shoreline of the Pennsylvanian sea had migrated northward to the vicinity of the town of Marathon and from there extended southeasterly along a line that roughly coincides with the Val Verde geosyncline.

In the structurally highest areas such as the Central Basin platform and the Marathon uplift the contact between the uppermost Pennsylvanian and the Wolfcamp of the Permian is quite evident and in places is even an angular unconformity. In these high areas there was obviously an interval of erosion between the latest Pennsylvanian and lowermost Wolfcamp. But in most sections of the Fort Stockton-Del Rio region and particularly along the Val

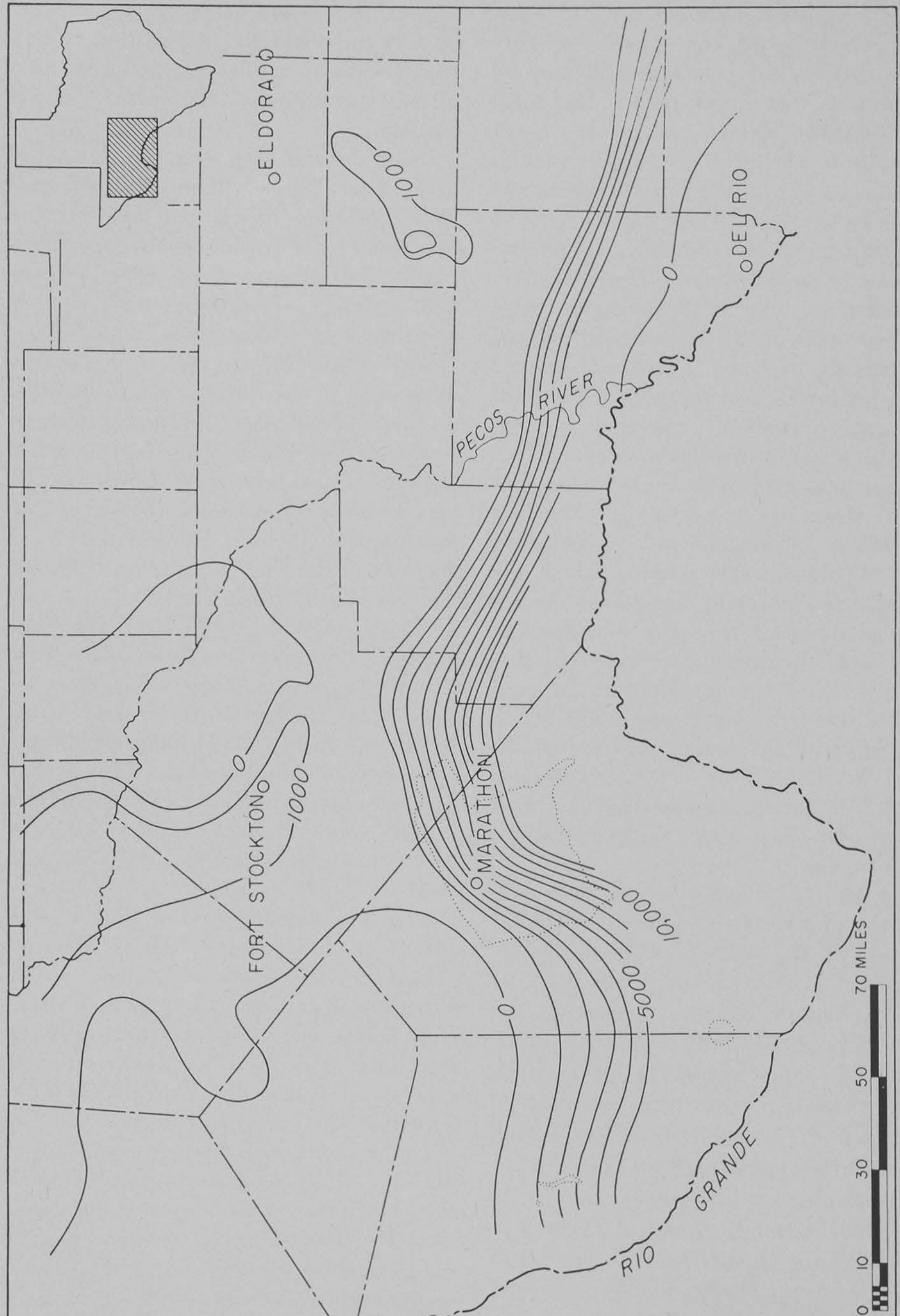



FIG. 60. Present thickness (in feet) of "lower" Pennsylvanian strata.


Fig. 61. Present thickness (in feet) of "upper" Pennsylvanian strata.

Verde geosyncline, there was no interruption in sedimentation and no obvious lithologic change, and it is a problem to separate Pennsylvanian from Permian within a generally uniform sequence of basin clastics. Fossils are rare but definitely show that in the geosyncline most of the thick section above lower Strawn limestone is of Wolfcamp age. One of the most clear-cut fossil finds turned up in Gulf Oil Corporation No. 1 P. G. Northrup, a deep test in eastern Reeves County in the deeper part of the geosyncline. Here a core taken about 1,000 feet above the lower Strawn yielded fusulines of lower Hueco, Wolfcamp, age.

Although there was no interruption of sedimentation at the Pennsylvanian-Permian boundary in the Val Verde geosyncline, there was from the beginning of Wolfcamp time a very great acceleration in the rate of downwarping of the geosyncline. This is apparent from the thickness map of the Wolfcamp (fig. 62), which reveals a maximum thickness of at least 14,000 feet. This enormous thickness has been revealed entirely by deep drilling within the past few years and previously was not even suspected. The type section of the Wolfcamp lies in the north edge of the Marathon Basin and at the southern margin of the Val Verde geosyncline; it has a measured thickness of 600 feet (King, 1937, p. 94).

The picture of the Wolfcamp is at present incomplete. A large segment of the south flank of the geosyncline in Terrell and Val Verde counties remains a blank on the map because there is simply no subsurface information. However, the available data do reveal the general shape, orientation, and linear character of the geosyncline. Also shown, but less precisely, is the branch which extends southwesterly across northwestern Brewster County and southern Presidio County and commonly is called the Marfa basin.

The quantity of sedimentary material shown by the isopach map is enormous. Excluding the Marfa basin and small parts of the geosyncline beyond the map area, the Wolfcamp contains 12,850 cubic miles of rock, and yet this deposit could have been built up at a rate of 1 inch every 50 years. The source of these sediments must have been equally large; 400 mountains the same size as the Franklin Mountains would have been needed to furnish the necessary rock debris.

These thick Wolfcamp geosynclinal deposits consist largely of interbedded shale and sandstone. The shale is dark gray and brownish gray, fine grained, and well consolidated. The sandstones are gray and brown, well cemented, commonly argillaceous, and generally fine to very fine grained. Some limestone is present in the upper Wolfcamp, and on the platforms, where the Wolfcamp is thin, this is the predominant rock type. On the Central Basin platform dolomite as well as limestone occurs in the Wolfcamp. This summary of the lithologic constituents of the Wolfcamp applies generally to the entire deposit, but the relative proportion of each lithic type varies from locality to locality. Figure 63 is a very generalized gross facies map of the entire Wolfcamp. The areas in which sand, shale, and limestone, respectively, are predominant are shown by distinct patterns. The largest area is covered by the predominantly shale facies; limestone is clearly associated with the platforms; and sandstone is the major constituent in a belt along the southern margin, including the Marfa basin.

This distribution pattern definitely suggests that most of the clastic constituents of the Wolfcamp of the Val Verde geosyncline came from the south. In that area there must have been an uplift of the pre-Wolfcamp formations of a magnitude sufficient to account for the thick Wolfcamp section. And since most of that uplift was composed of Pennsylvanian clastics, the lithologic similarity between Pennsylvanian and Wolfcamp is readily understood.

Not all of the Wolfcamp clastics were derived from the south. Locally in the deepest part of the geosyncline immediately south of the Fort Stockton high a few very deep tests have encountered lenses of detrital limestone in the lower Wolfcamp. These

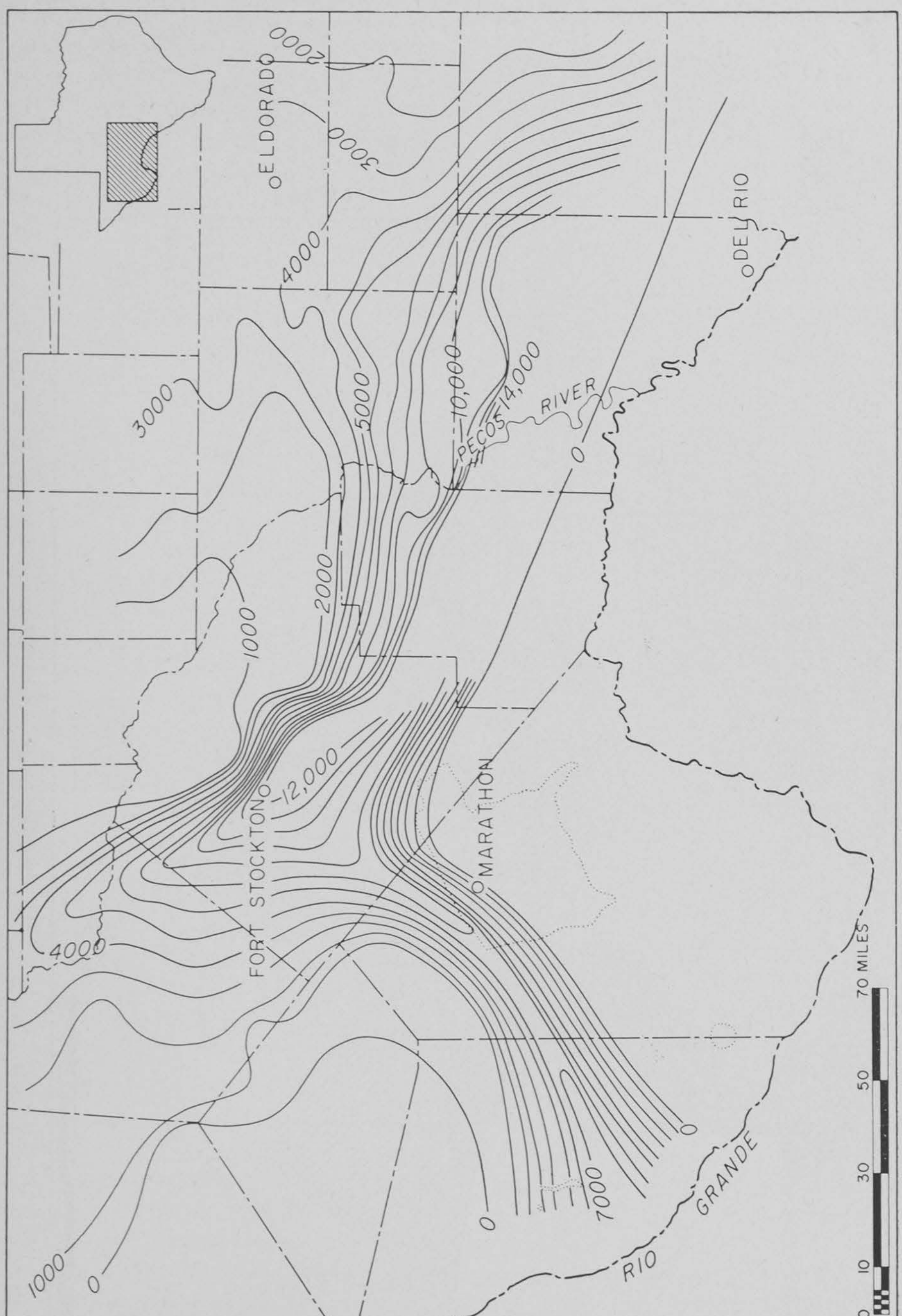



Fig. 62. Present thickness (in feet) of Wolfcamp series.

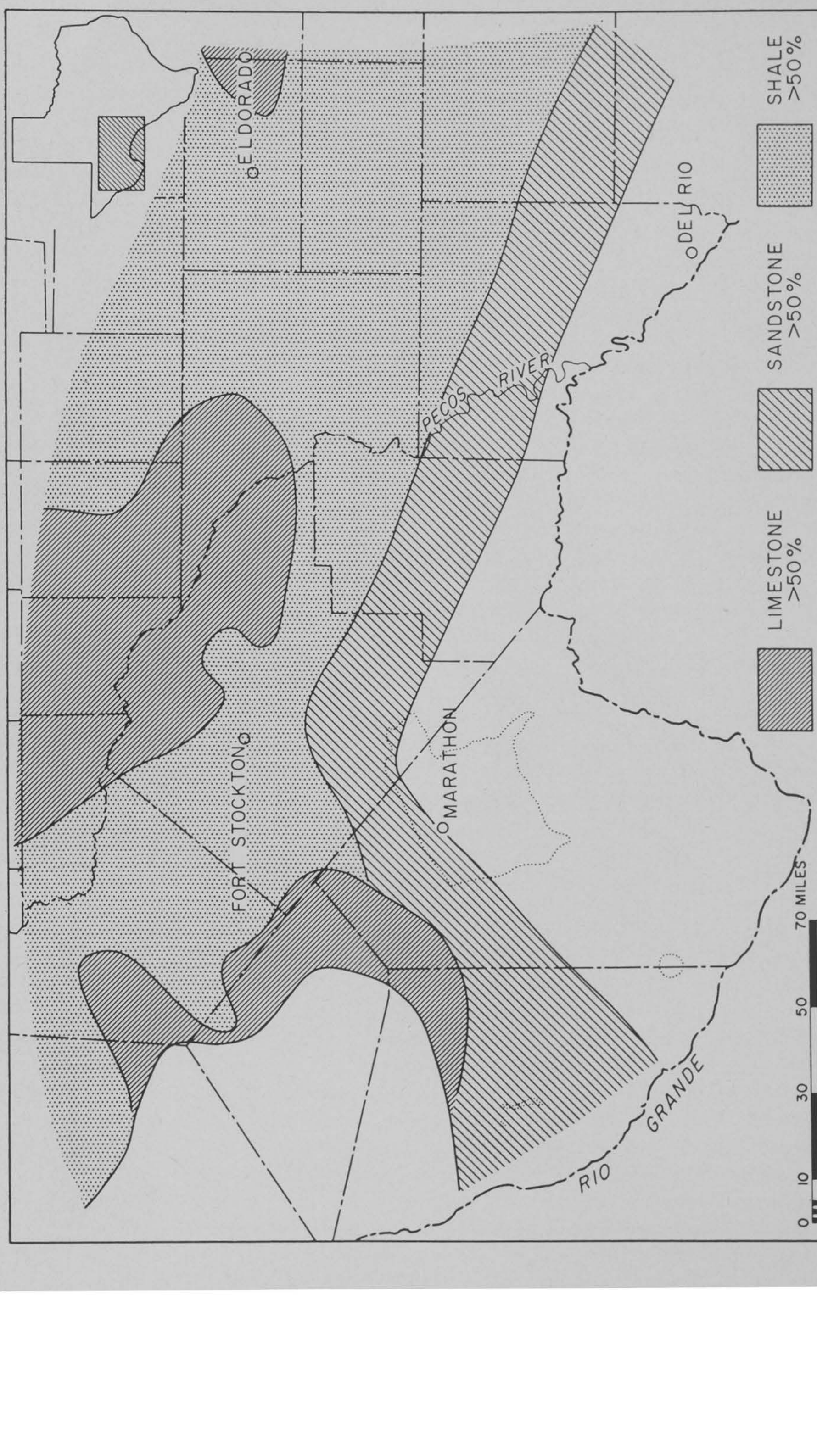



FIG. 63. Lithofacies of Wolfcamp sediments.

limestones have yielded Strawn and Canyon fusulines along with Wolfcamp species, and the implication is that these limestone fragments were eroded from Pennsylvanian reefs to the north on the flank of the Fort Stockton high. Obviously the reefs, as well as the crest of the high, must have been elevated during early Wolfcamp time.

A minor local facies of the Wolfcamp occurs well up on the south flank of the Fort Stockton high. Here the Precambrian igneous basement was exposed during a part of Wolfcamp time, and the lowermost beds of the Wolfcamp in the adjacent area to the south contain red shale and arkosic sandstone, apparently derived from the igneous exposure.

The absence of widespread marker beds within the Wolfcamp makes it difficult to work out the detailed sedimentary history of these deposits. Correlations may be carried limited distances and from these it appears that the older Wolfcamp strata are more restricted laterally than the younger units. This relationship is demonstrated, for example, by the above-mentioned occurrences of Pennsylvanian reef detritus in the lower Wolfcamp, because the same reefs were later overlain by a thick section of younger Wolfcamp beds.

The shape of these Wolfcamp sediments and their relation to older formations are illustrated in two stratigraphic cross sections (figs. 64, 65A). In both of these sections the upper horizontal line represents the top of the Wolfcamp or, where that is absent, the top of the next older unit present. The sections are drawn with a nearly five-to-one exaggeration of the vertical scale.

Figure 64 is located near the eastern end of the Val Verde geosyncline and extends a distance of 150 miles from a well near Del Rio to the western extremity of the Central Mineral region. The very great thickness of the Wolfcamp section as compared to all other Paleozoic units is immediately apparent; and within the Wolfcamp dashed correlation lines suggest the transgressive overlap of younger Wolfcamp

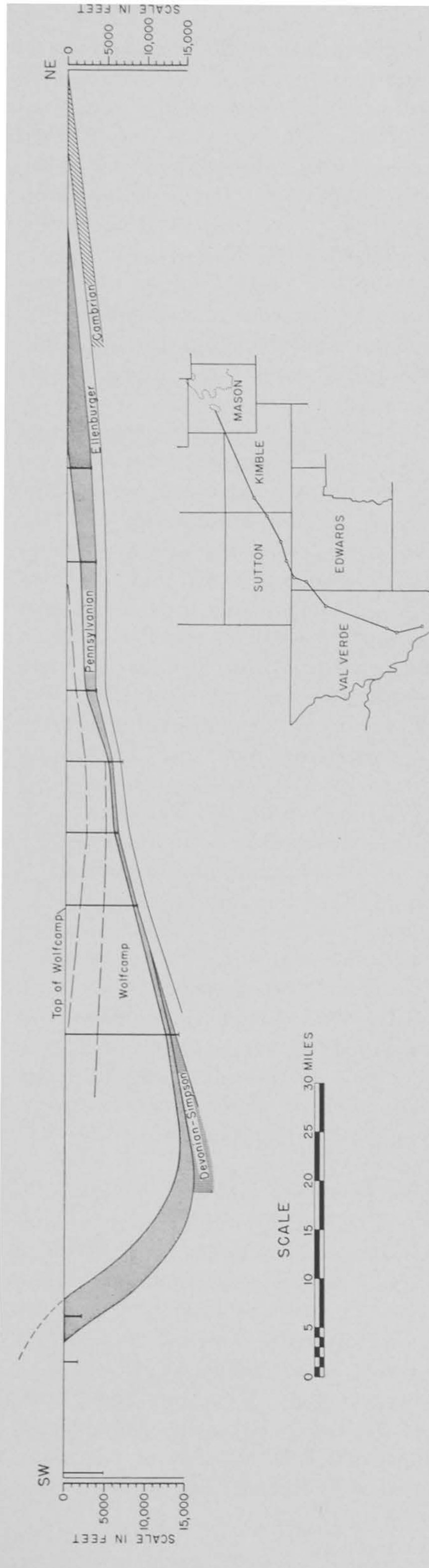



FIG. 64. Stratigraphic cross section through eastern Val Verde geosyncline.

beds beyond the older Wolfcamp strata. At the southwestern end of the cross section are indicated two wells which encountered no Wolfcamp but entered metamorphosed sedimentary rocks of probable lower Paleozoic age within the Devils River uplift (Galley, 1958). The manner in which this uplift meets the south limb of the Val Verde geosyncline is at present unknown because of a complete absence of well control and may be quite different from the uncomplicated north-dipping slope drawn on the cross section.

The second stratigraphic cross section (fig. 65A) is drawn southwest-northeast across the central Val Verde geosyncline from Gulf Oil Corporation No. 1 D.S.C. Coombs et al., a deep test near the town of Marathon, to Stanolind Oil and Gas Company No. 1 Conry-Davis Unit, a test near Horsehead Crossing on the Pecos River. Although scarcely over 90 miles in length this section includes parts of three geological provinces: the Marathon uplift, the Val Verde geosyncline, and the Central Basin platform. At the southwestern end of the section within the Marathon uplift the Gulf test was drilled through the Dugout Creek overthrust. This overthrust is shown on the section, but the other complicated structural features of that folded belt have been omitted. Along the line of figure 65A the Val Verde geosyncline has been crossed near its narrowest point, and several very deep tests serve to outline its general contours fairly well. However, along its south edge the relationship between the geosyncline and the Marathon uplift is not yet clear.

Three very deep wildcat tests have been drilled along the line of this section and are identified by the letters "A," "B," and "C." Well "A," Phillips Petroleum Company and Sinclair Oil and Gas Company No. 1-A J. C. Montgomery, reached a depth of 23,400 feet, having drilled 600 feet into the Ellenburger group. It encountered 12,000 feet of Wolfcamp sediments and appears to be located near the axis of maximum thickness of Wolfcamp beds. Well "B," Pan American Petroleum Company No. 1-CS University, was drilled to 21,687 feet, at which depth it was in shale of Mississippian age.

Well "C," Phillips Petroleum Company No. 1-EE University, which reached a depth of 25,340 feet (4.7992 miles), is [1959] the deepest test ever drilled into the earth's crust. Beyond this phenomenal depth, the well is also of very great interest because of unusual structural conditions which it encountered. The well penetrated a normal sequence of Cretaceous, Permian (including 4,800 feet of Wolfcamp), Strawn, and older Paleozoic formations to a depth of 13,765 feet, at which level a reverse fault was encountered, with Simpson overlying Devonian rocks. From that depth to 21,810 feet a whole series of structural abnormalities was met, and some units were repeated as many as four times. At the heart of this zone was a section of Ellenburger strata overlain and underlain by Simpson beds, the Ellenburger having an apparent thickness of 3,800 feet, although its true thickness is probably about 1,500 feet. The dipmeter recorded high-angle dips up to a maximum of 67 degrees. Below 21,800 feet no faults or other structural abnormalities were found; a normal sequence of lower Paleozoic strata was drilled, and the bottom of the hole is at a stratigraphic level 370 feet below the top of the Ellenburger group. The writer's interpretation, which is illustrated schematically in figure 65A, is that No. 1-EE University is located in a structurally complex zone of multiple faulting, including high-angle reverse faults and possibly some overturning, which separates the Fort Stockton high from the Val Verde geosyncline. A relative uplift of the Fort Stockton high of about 20,000 feet is shown. It appears that the well completely penetrated the disturbed zone, and the bottom 3,500 feet is in the relatively undisturbed segment of the crust which forms the deeper portion of the Val Verde geosyncline.

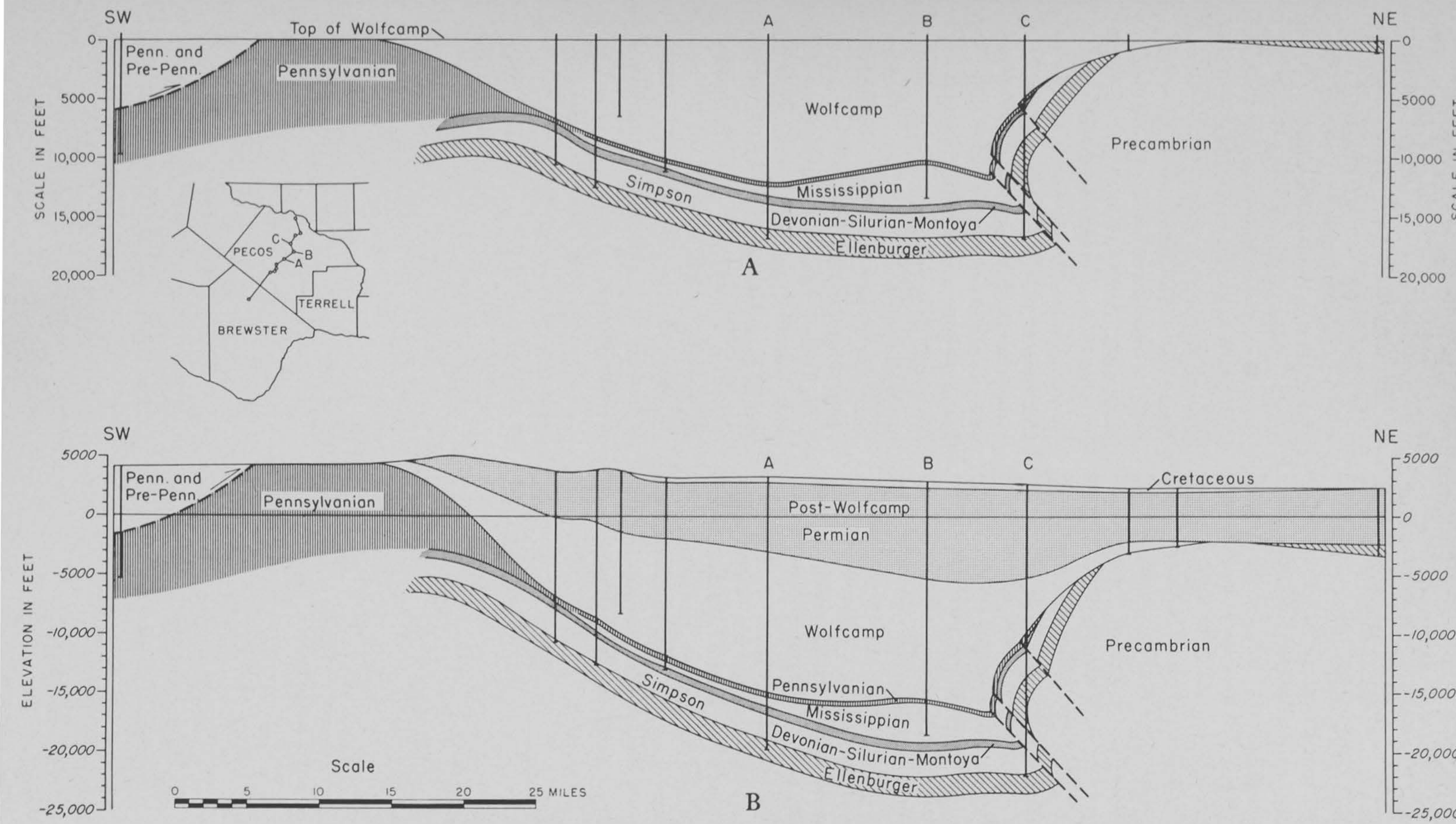



FIG. 65. Sections across central Val Verde geosyncline. A, Attitude of formations prior to Leonard time. B, Present attitude of formations.

## Post-Wolfcamp History

Nearly everywhere throughout the Fort Stockton-Del Rio region the close of the Wolfcamp was a time of structural quiescence, and sedimentation proceeded without interruption into Leonard time. Consequently, there is in most areas no lithologic break at the top of the Wolfcamp, and the horizon is difficult to identify except where fossils happen to appear in samples or cores. However, the Leonard sea was not as extensive as in Wolfcamp time; the sea withdrew from the southeastern segment of the Val Verde geosyncline and remained only over the portion which lies northwest of northwestern Val Verde County. In that area were deposited the platform carbonates, basin shales, and argillaceous limestones of Leonard time, and the still younger siltstones, sandstones, dolomite reefs, and evaporites of Guadalupe and Ochoa times. The area of maximum sedimentation was the Delaware basin, mostly north and west of Fort Stockton. A narrow extension of this basin ran west-east across Pecos County just south of the Fort Stockton high and served as a connecting depression between the Delaware and Midland basins. This synclinal feature has long been known as the Sheffield channel (Cannon and Cannon, 1932, p. 199); it is not to be confused with the earlier and much larger Val Verde geosyncline. These post-Wolfcamp sediments have been described and will not be reviewed in this brief paper (Galley, 1958; Adams, 1944).

Figure 65B is a diagrammatic cross section which is drawn to the same scale and through the same points as the section of figure 65A but differs from that section in being a representation of present-day geology and, therefore, includes all strata. All post-Wolfcamp Permian beds are included in a single unit which is considerably thinner than the Wolfcamp alone. By comparison with figure 65A the Wolfcamp section exhibits nearly the same shape but has undergone a mild tilting down toward the north. Present structure of the base of the Wolfcamp is given in figure 66, which bears a very close resemblance to the Wolfcamp thickness map (fig. 62).

No Triassic beds occur on the line of the cross section (fig. 65B), although a thin unit of these continental red beds is generally present over the nearby Delaware basin and Sheffield channel. At the top of the section a very thin layer of Cretaceous sandstone and limestone completes the geologic picture.

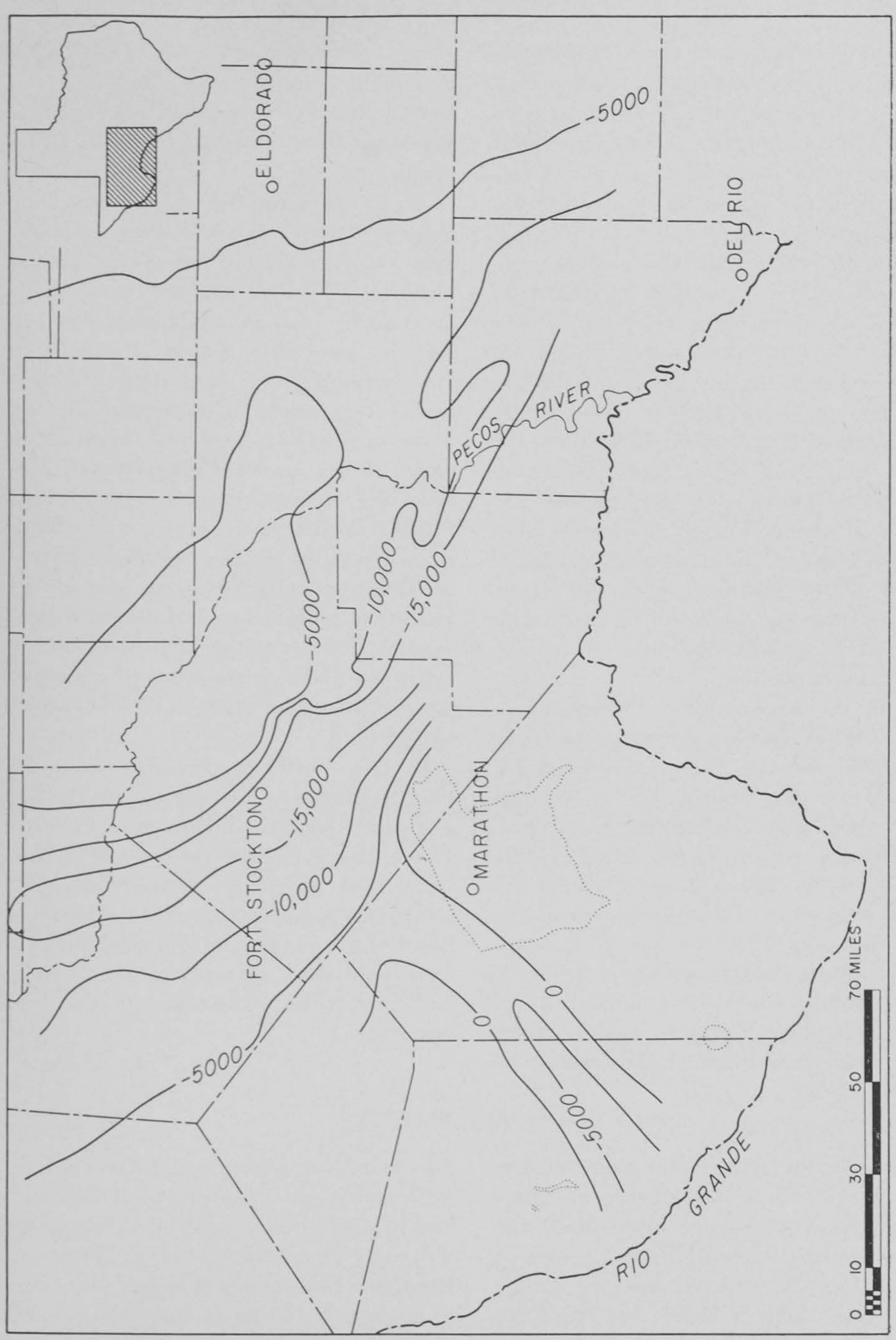



FIG. 66. Structure of base of Wolfcamp series.

## SUMMARY

The Paleozoic history of the Fort Stockton-Del Rio region began with peneplanation of an ancient terrane followed by a late Cambrian northward transgression of the sea across Texas and adjacent areas. Thereafter, the region was only part of a much larger sea which endured throughout most of Paleozoic time. During the Simpson to Woodford interval of the Paleozoic, the sea was somewhat more restricted than either before or after and occupied the shallow Tobosa basin; the Fort Stockton-Del Rio region comprised the southern half, or less, of that basin. Throughout the great length of time from Cambrian through Mississippian, the region was relatively stable; crustal movements were regional but mild. Sediments deposited under these conditions were thin; they were largely carbonates until the beginning of Woodford time and, thereafter, primarily dark shale.

Near the beginning of Pennsylvanian time there occurred a profound change in structural activity. A major geosyncline formed near the southern border of Texas and extended far beyond the Fort Stockton-Del Rio region. It was filled rapidly with a thick accumulation of clastic deposits derived from a contemporaneous uplift to the south. Still in "lower" Pennsylvanian time the geosynclinal sediments were greatly folded and faulted in an Appalachian-type orogeny, and at about the same time the major uplifts and basins of the foreland area of west Texas came into being. At this time the Val Verde geosyncline first appeared as a definite structural feature, although without great depth.

About the beginning of Wolfcamp time orogenic activity recurred along the Marathon belt, and during this epoch the Val Verde geosyncline attained most of its great depth. In it accumulated at least 14,-000 feet of clastics, mainly derived from Pennsylvanian and older rocks in highlands immediately to the south. The Val Verde geosyncline developed considerably north of the earlier Pennsylvanian geosyncline, as was pointed out by Hall (1956). Well after the beginning of Wolfcamp deposition a major overthrust carried pre-Permian sediments over part of the Val Verde geosyncline, and this movement may be considered the final phase of the major late Paleozoic orogeny which began many miles to the south in early Pennsylvanian time.

In post-Wolfcamp Permian time the Fort Stockton-Del Rio region returned to a state of relatively mild crustal activity. The southeastern part was probably a low-lying land area; the northwestern part formed the southern corner of the Permian basin and was the site of deposition of a moderately thick sequence of typical basin and platform carbonates, clastics, and evaporites.

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## REFERENCES

ADAMS, J. E. (1944) Upper Permian Ochoa series of Delaware basin, west Texas and southeastern New Mexico: Bull. Amer. Assoc. Petrol. Geol., vol. 28, pp. 1596–1625.

————, FRENZEL, H. N., RHODES, M. L., and JOHNSON, D. P. (1951) Starved Pennsylvanian Midland basin: Bull. Amer. Assoc. Petrol. Geol., vol. 35, pp. 2600–2607.

CANNON, R. L., and CANNON, JOE (1932) Structural and stratigraphic development of south Permian basin, west Texas: Bull. Amer. Assoc. Petrol. Geol., vol. 16, pp. 189–204.

CHENEY, M. G. (1929) History of the Carboniferous sediments of the Mid-Continent oil field: Bull. Amer. Assoc. Petrol. Geol., vol. 13, pp. 557–594.

GALLEY, J. E. (1958) Oil and geology in the Permian basin of Texas and New Mexico, *in* Habitat of oil, Amer. Assoc. Petrol. Geol., pp. 395–446.

HALL, W. E. (1956) Marathon folded belt, Big Bend, Texas: Bull. Amer. Assoc. Petrol. Geol., vol. 40, pp. 2247–2255.

KING, P. B. (1937) Geology of the Marathon region, Texas: U.S. Geol. Survey Prof. Paper 187, 148 pp.

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